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ASPECTS OF EVAPORATION AND EVAPOTRANSPIRATION IN THE WATER BALANCE OF BAKER CREEK BASIN, NEAR YELLOWKNIFE, NORTHWEST TERRITORIES

by

© JOHN BRADFORD WIGHT

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF MASTER OF SCIENCE

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA
FALL, 1973

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled ASPECTS OF EVAPORATION AND EVAPOTRANSPIRATION IN THE WATER BALANCE OF BAKER CREEK BASIN, NEAR YELLOWKNIFE, NORTHWEST TERRITORIES submitted by JOHN BRADFORD WIGHT in partial fulfilment of the requirements for the degree of Master of Science.

Arleigh H. Laycock
Supervisor

William C. Wonders

Lawrence C. Bliss

Date Sept. 27, 1973

ABSTRACT

This study is an attempt to define more closely the average annual patterns of evaporation and evapotranspiration for surfaces in the Baker Creek Basin, Yellowknife, N.W.T.- a small subarctic research basin located in the discontinuous permafrost zone and within the Canadian Shield. Specific emphases in the study are the elucidation of patterns of evaporation and evapotranspiration for certain vegetation type found in the basin, and of lake evaporation patterns. Perspective on, and partial verification of, these patterns was obtained by an assessment of the average annual basin-scale water balance.

The research comprises: a comprehensive literature review; net radiation, evaporimeter, soil moisture and active layer studies at muskeg, sedge meadow and Black Spruce woodland bog sites; lake surface temperature surveys for formulaic estimations of lake evaporation; analysis of published lake level and flow records for the Pocket Lake sub-basin; analysis of published runoff data for Baker Creek; and a synthesis of the average annual basin water balance after three approaches - Thornthwaite water balance model, climatonomic analysis, and semi-empirical studies. With this synthesis, refined average annual water balance equations were suggested for various surfaces in the basin.

It was concluded that the process of evapotranspiration from vegetative surfaces did not function efficiently at the study sites in terms of the available radiant energy and

soil moisture considerations. Suggested reasons for this effect were the non-vascular nature and mulching properties of the moss-covered sites or the inhibition of transpiration through the vascular flora. The latter possibility was thought to be related to distinctive edaphic factors associated with the permafrost where it was present. After certain assumptions about evaporation from bedrock-dominated surfaces were made it was estimated that evapotranspiration from land surfaces in the basin averaged seven inches.

With further interpretation it was concluded that the magnitude of the lake evaporation parameter may have been underestimated in the past, possibly due to hitherto unappreciated local advection effects or by previously exaggerated heat storage lag effects and early season condensation gains. It was estimated that free water evaporation over the basin averaged eighteen inches which is two inches greater than the evaporating power of the effective annual net radiation and four inches greater than certain of the mean small lake evaporation estimates published in the literature.

Basin precipitation estimates derived from Yellowknife Airport records were deduced to require only an upward adjustment of about ten per cent for them to be acceptable as being probably representative for the basin.

The average annual basin water balance (1968-1972) was deduced to be:

Precipitation = Runoff + Evapotranspiration

(inches) 11.0 = 1.4 + 9.6

The level of Potential Evapotranspiration was considered to be close to the sixteen inches generally quoted for the area. This is well above the annual precipitation and the runoff from the basin must be interpreted as being largely the product of exceptional precipitation events and particular surface storage relationships. This runoff would be greater if depression and lake storage was not such an important feature of the basin hydrology.

ACKNOWLEDGEMENTS

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CHAPTER I

INTRODUCTION

In this study an attempt is made to define more closely the patterns of lake evaporation and evapotranspiration from land surfaces in the Baker Creek Basin, near Yellowknife, N.W.T. This is a small research basin located in the subarctic climatic zone, the discontinuous permafrost zone and the Canadian Shield physiographic province.

The headwater areas of Baker Creek are located about fifteen miles north of Yellowknife, N.W.T., and the creek enters Yellowknife Bay on the North Arm of Great Slave Lake about two miles north of Yellowknife. The basin lies between 62°28' and 62°43' North latitude and 114°17' to 114°31' West Longitude. Yellowknife is slightly more than 600 air miles (or 1000 highway miles) north of Edmonton, Alberta (Figure 1-1)

Baker Creek basin is one of a series of representative basins selected for special study under the International Hydrological Decade program (1965-1974) (File Reference: 1WB-RB-44; NORTH 9). The total basin area is approximately seventy square miles and the drainage area guaged is fifty-five square miles (Baker Creek Main Stem: Station Number 07SB009). The basin research is administered by Professor A.H. Laycock, Department of Geography, University of Alberta, who selected the basin for its good northern representativeness, its relative accessibility by air and road, and the

FIGURE 1-1.
LOCATION OF BAKER CREEK WITHIN THE MACKENZIE RIVER DRAINAGE BASIN

Source: Landals,(1970,p.xvi)

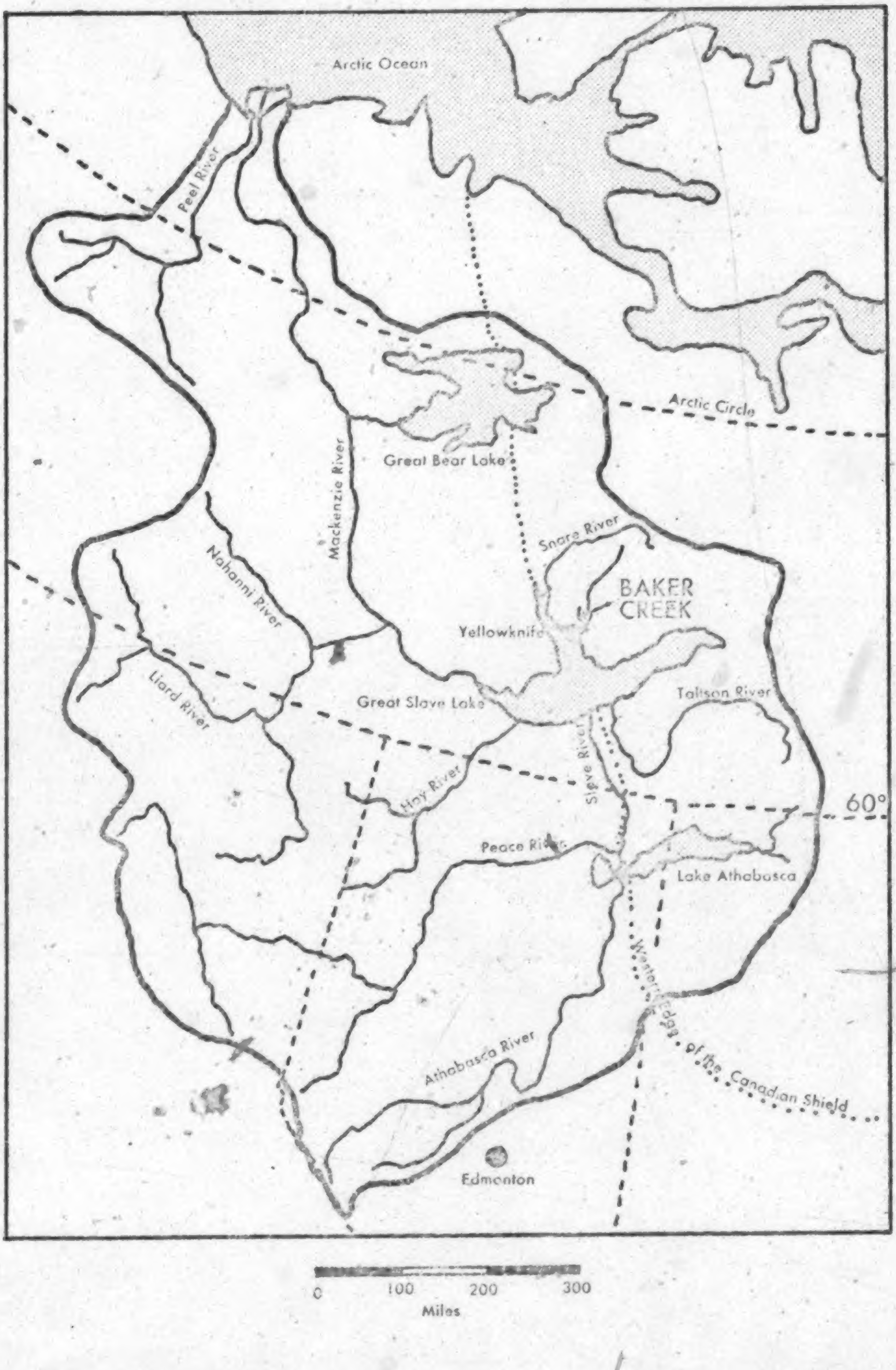


FIGURE 1-2. BAKER CREEK BASIN AND LOCATIONS OF ASSOCIATED METEOROLOGICAL AND HYDROLOGICAL STATIONS.

Source: Landals,(1970,p.xv)

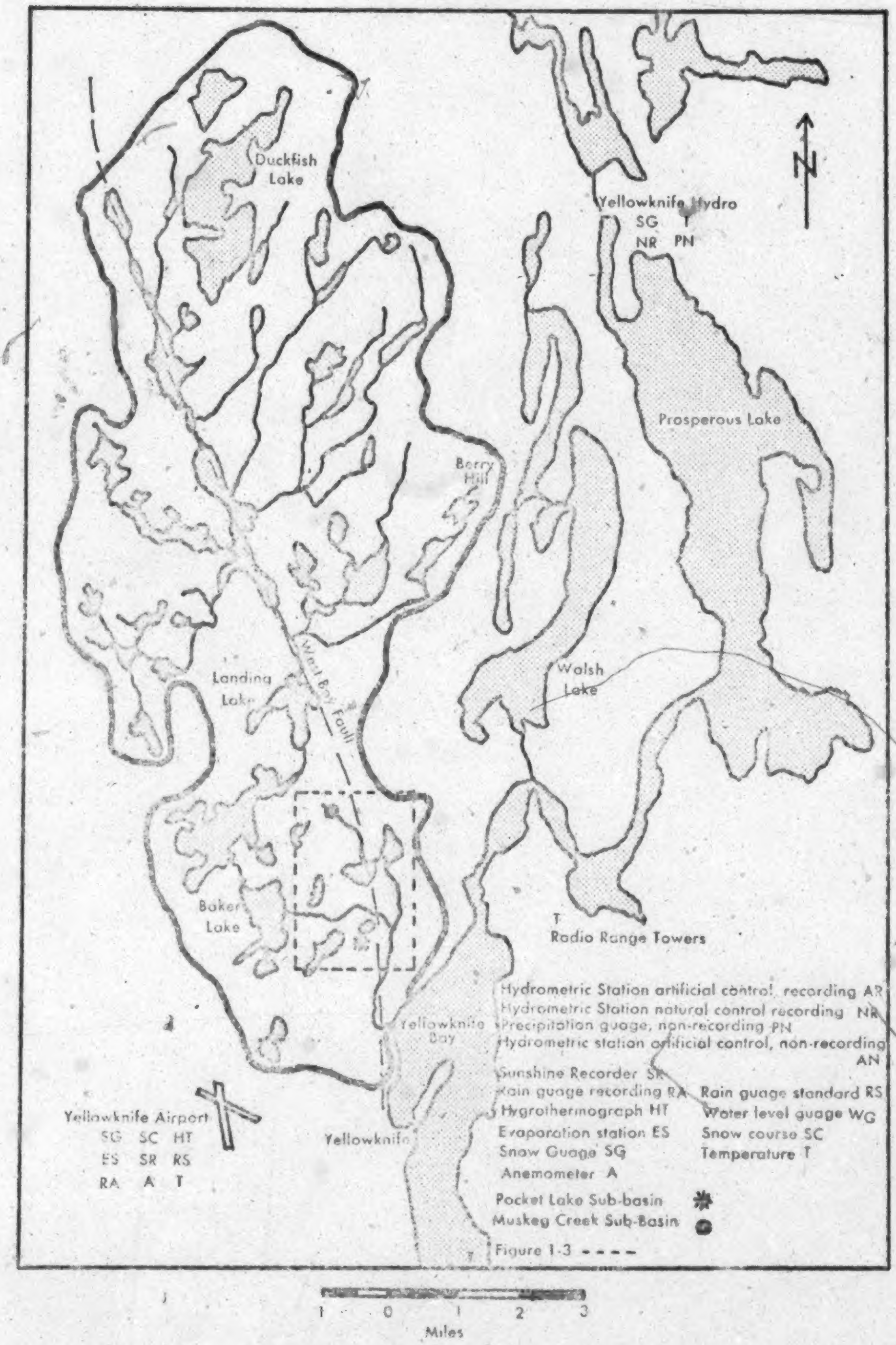
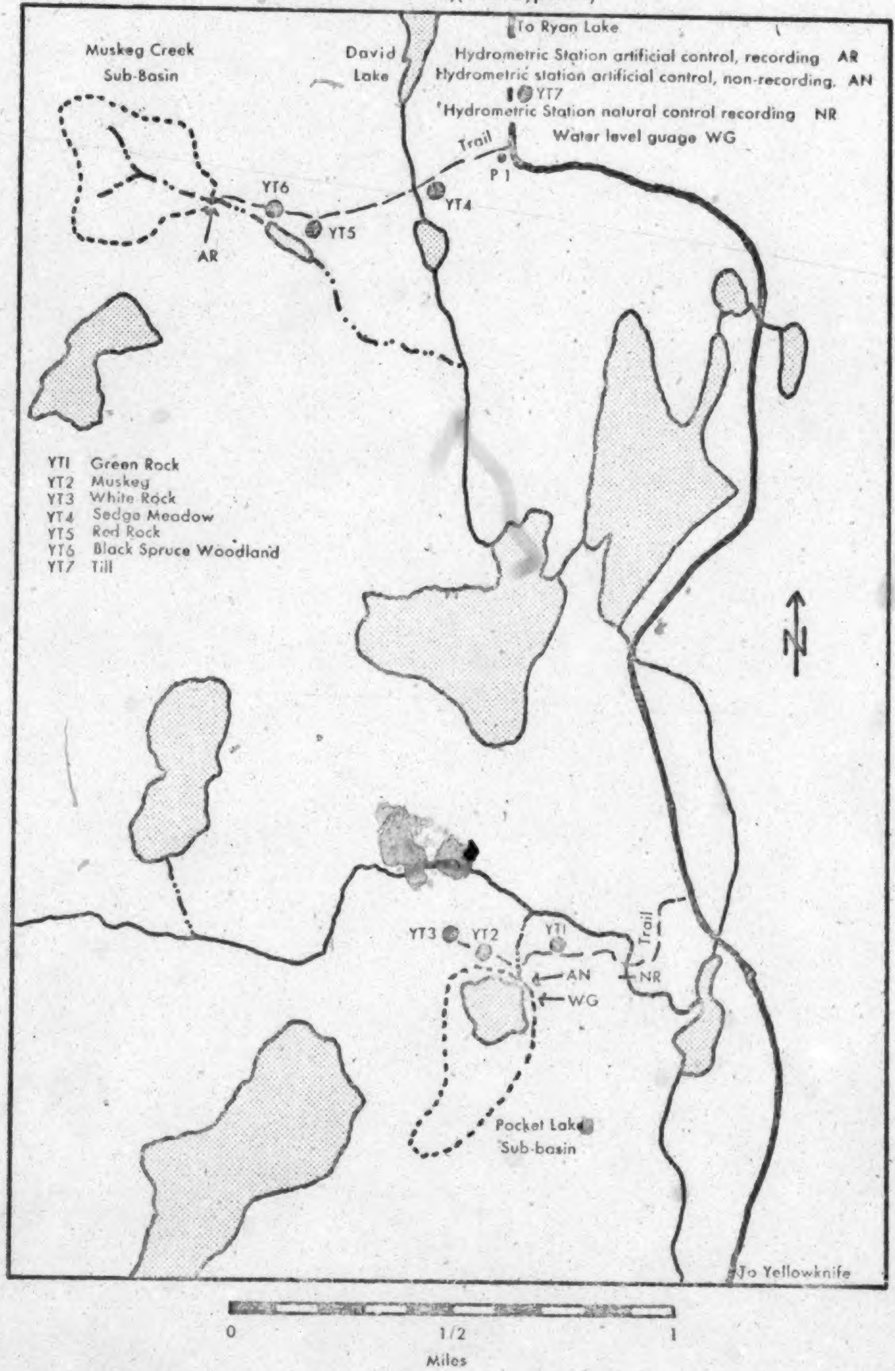


FIGURE 1-3 MUSKEG CREEK AND POCKET LAKE SUB-BASINS SHOWING HYDROLOGICAL STATIONS, GROUND TEMPERATURE SITES AND RYAN LAKE ROAD.

Source: Landals,(1970,p.xvi)



proximity of stations with a reasonable record of meteorological and hydrological data (Figure 1-2).

Pocket Lake and Muskeg Creek are two sub-basins (Figure 1-3). The hydrometric station at Muskeg Creek has not functioned satisfactorily owing to leakage problems caused by permafrost active layer movement. The lake level recorder at Pocket Lake has functioned very well and the record has proven very useful in estimating lake evaporation.

Much of the hydrological work in Baker Creek basin has been undertaken in conjunction with ground temperature measurement studies undertaken by R.J.E. Brown to determine aspects of the distribution of permafrost in the region. Seven of the eight jointly developed sites in the Yellowknife area occur within the basin and the location of these are noted in Figure 1-3. Much fieldwork in the present study was undertaken at three of these sites--YT2, YT4, YT6.

Research conducted in the basin to date has been focussed upon snow and snowmelt runoff studies (Kakela, 1969; Landals 1970). The present research is complementary to the previous research in that its focus is upon the summer season and the patterns of evaporation and evapotranspiration. Funding for the present study was obtained from a N.R.C. grant received by Professor Laycock and a grant-in-aid from the Boreal Institute, University of Alberta.

RESEARCH PROBLEM

As is the case for many other northern locations the data presently available for Baker Creek basin are not satisfactory for close definition of the water balance patterns in the region. In the present study an attempt is made to resolve this problem with a focus on the nature of the evaporation and evapotranspiration patterns present in the study area. These latter patterns have suffered a particular lack of attention not only in Northern Canada generally but also in Baker Creek basin where previous research has focussed on winter and spring seasons when evapotranspiration and evaporation are of negligible importance in the water balance.

Evaporation has been called "the most desperate branch of that desperate science, (meteorology)" (Bruce and Maasland, 1969, p 69). The problems which confront its measurement and accurate estimation in southern regions (Hare, 1961) are compounded by the under-researched influences of several distinctly northern characteristics. Most notable among these characteristics appears to be a lack of knowledge of the pattern of water loss as a result of evaporation and evapotranspiration processes through the vegetation indigenous to the Subarctic. The few studies which have been made to date indicate that the evaporation process operates at significantly lower rates through the moss and lichen covers which dominate much of the region (Nebiker and Orvig 1958; Watts, Galloway and Grenier, 1960; Brown 1961, 1965). This evidence has been obtained mainly from evaporimeter and atmometer measurements. Corroborative evidence has been obtained by means of a heat balance approach. In northern areas such studies have indicated substantially lower rates of utilization of net radiation by evaporative heat flux by comparison with mid-latitude regions (Mather and Thornthwaite, 1956, 1958; Ahrnsbrak 1968; Rouse and Kershaw 1971, 1973). The combined evidence from such studies does not appear to be strong enough yet to justify projection of the results over all of the North. More studies of a similar nature appear to be required in other parts of the North before a clearer definition of this component's patterns can be attempted.

Much of Northern Canada is covered by lakes, particularly in the Canadian Shield region, and lake evaporation would appear to be a relatively significant component of the northern water balance. Any regional estimate of evaporation must include close consideration of lake evaporation. Little substantial research has been undertaken in the North however to function as a guide in making such estimations. The results of studies which have been made have generally been obtained by application, to scanty climatological data, of empirical formulae derived in mid-latitude regions (eg. Richards & Webb, 1971). More research is required to check the results obtained by such methods, particularly through an evaluation of the influence of the heat storage lags and local advection effects which might be expected to alter the evaporation regimes of northern lakes.

Regional evapotranspiration estimates can also be derived as a residual in the water balance equation. In this approach it is assumed however that precipitation, runoff and storage effects may be accurately estimated. In the North basic measurements of precipitation and runoff are more abundant than evapotranspiration estimates. In the striving for a better definition of the water balance these two components have received much attention, particularly precipitation (Hare and Hay, 1971; Findlay 1969). Improved estimates of the precipitation variable have permitted significant improvements to be made in the definition of the water balance. By contrast few attempts have been made to solve the problem by research designed to determine estimates of regional evapotranspiration for use in combination with more representative runoff measurements. This is however the broad thrust of the present thesis.

OBJECTIVES

In the initial research design a relatively narrow study focus upon net radiation patterns and evapotranspiration relationships for sample sites present in the basin was envisaged. Severe instrument problems arising in the course of field research prevented the satisfactory realisation of this objective and resulted in a need for a re-orientation of the thesis design along much broader lines. Thus the scale focus of the thesis shifted from an emphasis on the site scale to one on the basin scale. The basin scale conclusions were derived or inferred from analyses of fieldwork results, literature research and available standard climatological and hydrological data.

It may be convenient to list the revised objectives on the basis of scale.

Site scale: The measurement and/or estimation of evaporation of evapotranspiration from: certain understory vegetation types found within muskeg, sedge meadow and Black spruce woodland bog plant associations; a sample of lake surfaces; and some small free water surfaces. These objectives relate mainly to the study period June to August, 1972.

Surface scale: Estimation of average annual water balances for the main surface types present in the basin - vegetative, free water and bedrock-dominated surfaces. (The latter surface type is distinct from bare rock surfaces).

Basin scale: A more satisfactory definition of the average annual water balance patterns for Baker Creek basin with a focus upon the evapotranspiration component of the water balance system.

METHODS

The thesis is developed by building on conclusions derived at each scale of investigation from the sites to the basin. The detailed fieldwork research which could be salvaged from the initial research design was used at a site scale. The sites thus intensively investigated constituted a distinctly limited sample of the total variety present in the basin. Results from these investigations are intended to be viewed largely as sample checks on patterns deduced from analyses of literature research and available standard data. For the vegetative surfaces data were obtained mainly by evaporimeter studies in conjunction with soil moisture measurements. Net radiation was monitored almost continuously at one of the vegetative sites. More information, of a general nature, on heat balance aspects was obtained at the permafrost sites where active layer development was monitored by periodic probing of representative surface areas. The equilibrium evapotranspiration model (after Rouse and Stewart, 1972) was also used to obtain estimates of evapotranspiration at the permafrost sites.

Evaporation from certain lake sites for the summer of 1972 was estimated using lake surface temperature measurements in a semi-empirical formula (Richards and Webb, 1971). Lake level changes in a small lake-dominated sub-basin (Pocket Lake) were also utilised to derive lake evaporation estimates for the summers of 1967-1972. The changes in level of small pools during the summer of 1972 were also monitored in order that evaporation patterns at such sites might be estimated.

At the surface scale, generalizations from site scale data, derived primarily from fieldwork observations, were heavily supplemented by inferences drawn from an extensive literature review and the interpretation of available climatological and hydrological data. It was also at this scale that generalizations were made concerning the water balance patterns of the bedrock-dominated surfaces. This surface type constitutes approximately 50-60 percent of the basin area. (Bare rock surfaces would probably constitute a significantly smaller proportion of the basin area). Although no detailed fieldwork was undertaken on bedrock-dominated sites it was thought possible to generalise concerning the patterns present through an appreciation of the interrelationship of storage capacity and precipitation might be obtained. This appreciation was improved by reference to field observations at typical bedrock sites in the basin made by Kakela (1969) and Landals (1970).

Synthesis of the basin scale water balance was obtained primarily by application of the results of the research at the site and surface scale in calculations involving three separate approaches aimed at realising a better definition of the water balance of the basin. These three approaches were the Thornthwaite water balance (Thornthwaite and Mather 1955; 1957), climatonomic analysis (Lettau 1969; Hare and Hay 1971), and semi-empirical studies orientated to refining the work done by Kakela (1969). The implementation of these approaches was greatly facilitated by the use of good runoff records for the Baker Creek basin and Pocket Lake sub-basin for the years 1968-1972. These records therefore have received more attention in the overall analysis than may have been anticipated for a thesis with a focus upon evaporation and evapotranspiration.

A major outcome of the revised, broader thesis design was a greater emphasis on the literature review as an initial source of information for deriving water balance patterns in the basin. In several cases the material covered has been used to supply conclusions which could not be wholly substantiated from the fieldwork conducted during the present study. For instance, this approach was adopted in estimating the nature of evapotranspiration for the trees and shrubs which constitute an important part of the vegetation in the basin. Field methods to obtain good first-hand estimates for these parameters were not practical in terms of the necessary instrumentation and operator expertise.

The literature review has also been used to overcome the consequences attendant upon instrument failure and other fundamental measurement problems encountered during fieldwork. Extensive use was made of the literature on albedo patterns over subarctic surfaces to gauge the net radiation over different sites from a record for one site. This was made necessary by the failure of one portable and one recording net radiation measurement systems. In the initial research design these systems were to be periodically rotated around a selection of natural sites and related to the continuous control record. It was only possible to use the control record by itself however. Similarly the literature on ground heat flux patterns in subarctic and arctic areas was drawn upon extensively to replace unsatisfactory field estimates. This situation was precipitated by an apparent malfunctioning of the potentiometer used to measure the thermocouple temperatures. There were also difficulties in obtaining representative estimates of the thermal conductivity of the organic and mineral soils monitored.

ORGANIZATION OF THE STUDY

The physical setting of the study area as it relates to the objectives of the study is described in Chapter Two. Chapter Three is a comprehensive review of the pertinent literature. In Chapter Four the intensive study sites, research methods employed, and limitations of the study's findings are described. Chapter Five contains background climatological information to provide a perspective on conditions during the study period in relation to long term patterns. This chapter is mainly a presentation of the results gained from the research into evaporation and evapotranspiration patterns for various surfaces,and an interpretation of these results. In Chapter Six an attempt is made to synthesise a better definition of the water balance of Baker Creek basin. Mean annual water balance equations are presented for various sites in the basin and compared with those determined by Kakela (1969). Climatonomic analysis is also employed to determine the best combination of parameters necessary for more precise definition of the mean annual water balance of the basin. Chapter Seven contains a summary of the research, conclusions drawn from it and suggestions for future research.

CHAPTER TWO

THE PHYSICAL SETTING

INTRODUCTION: The following description of the physical setting of the Baker Creek basin will provide the perspective necessary for proper evaluation of this study's results. In their theses, Kakela (1969) and Landals (1970), accomplished this descriptive task with lengthy chapters. The systematic survey presented here is in part a summary of their work. This presentation however contains an emphasis on those aspects which particularly influence the summer water balance of the region.

BEDROCK GEOLOGY

Baker Creek basin is situated within the Canadian Shield as opposed to most of the Mackenzie River drainage basin which is located immediately over bedrock of sedimentary origin. The complex geological history of the region has resulted in the outcropping of a number of contrasting rock types. These distinctive lithologies, because of differences in weathering patterns, porosity differences and differences in albedo affect evapotranspiration patterns through their more direct influence on runoff patterns.

About 85 percent of Baker Creek basin is underlain by Archaean granodiorite, granite and allied intrusive rocks. The remainder of the basin is mainly within the Yellowknife Greenstone Belt which is located about the centre of the eastern edge of the basin. A few large basic dyke systems also occur in the basin.

Faulting is very prominent. The trend of the faults is north and northwest with some large cumulative horizontal and vertical displacements. The West Bay Fault which divides the basin roughly in half from northwest to southeast is the largest fault in the region and has been traced for more than 150 miles. Movement along this fault has resulted in a definite lineation of lakes in the basin and to a large extent the fault controls the drainage pattern of Baker Creek. From the northern end of the basin, southward to Landing Lake, the course of Baker Creek itself is strongly controlled by the West Bay Fault. Tributaries generally flow southwest toward this fault and in many cases are controlled by small cross faults, dykes and veins.

Local relief seldom exceeds 200 feet and Berry Hill, on the eastern divide, approximately 880 feet above sea level and 370 feet above the level of Great Slave Lake, is the highest point in the basin. Granitic areas have a somewhat different physiography than the greenstone areas of the Yellowknife Group. According to Jolliffe (1945) areas underlain by granitic rocks show the least local relief and commonly stand at a higher average elevation than areas underlain by rocks of the Yellowknife Group.

SURFICIAL DEPOSITS

Most of the surficial deposits in the area are materials deposited directly from glacial ice or by glaciofluvial and glaciolacustrine action. On the basis of stratigraphical relationships and radiocarbon dating it has been suggested by

Craig (1964, p 39) that the area just east of the basin was deglaciated about 7000 years B.P.

The primary result of the glaciation was the reduction of local relief. Glacial erosion scoured and rounded the upland areas while deposits partly filled the adjoining depressions. Roche moutonées are common features in the basin as are striations, chattermarks and crescentic gouges. The orientation of these features indicates an ice movement direction from the east-north-east. Glacial erosion has also produced numerous hollows of various size which act as semi-permanent reservoirs to reduce or detain runoff and increase evaporation.

Glacial erosion on a microscale can be very significant in terms of evapotranspiration losses because of its effect on detention capacities and infiltration rates. Although many of the bedrock outcrops appear relatively smooth from a distance, on closer inspection many roughened pavement areas and small scale depression areas can be observed (Plates 2-1 and 2-2). Some have trapped windblown soil material and vegetation has developed. Because of these features detention storage is increased and actual runoff is less than what might be expected from such apparently smooth and bare bedrock areas. This pattern is accentuated by the nature of the precipitation events since the small depressions act to detain for evapotranspiration most of the precipitation of lighter rains. Only a few heavy rains, and the snow-melt surplus, have significant yields on such surfaces.

Plate 2-1. Roughened granite bedrock pavement with relatively large detention capacities. Note mosses in depressions where trees can take root. Photograph taken on the edge of Muskeg Creek sub-basin facing in a southwesterly direction.

Plate 2-2. Greenstone bedrock near Ryan Lake road. Relatively smooth rock surfaces interspersed with numerous depressions containing an absorbent vegetation cover. Photograph taken facing in a northeasterly direction.

Deposition of glacial till has not been of marked significance in Baker Creek basin. According to Jolliffe's map (1945) only about 5 percent of the basin is covered by 'major' areas of glacial till. The distribution of this type of deposit is mainly restricted to the southeastern divide region of the basin. Major glaciofluvial deposition areas within the basin appear to be similarly distributed as far as the basin is concerned although some deposits may be more widespread but concealed under peat covers in depressions.

Another category of glacial melt water deposits are those associated with proglacial lakes. These materials are of two main types: fine lacustrine sediments and beach sands and gravels. The more extensive of these deposits may be interpreted as being products of Glacial Lake McConnell which has been identified as extending from Great Bear Lake, through Great Slave Lake to Lake Athabasca (Craig 1965). This extensive proglacial lake was formed immediately after deglaciation. Beach deposits associated with it are locally very prominent and Jolliffe (1945) identified these features in Baker Creek basin up to elevations of 240 feet above the present level of Great Slave Lake. These beach deposits consist of well-sorted sand and fine gravel, are generally well-drained, and provide optimal locations for vegetation growth.

The finer glacio-lacustrine deposits are also important locally though they are not as extensive as the beach materials. The proglacial activity resulted in the movement of material to the bases of depressions and the washing of the slopes. The resulting contrast between cleanly-washed bare rock slopes and lacustro-till filled depressions is greater than in areas not subject to lake action. The lacustrine deposits consist mainly of fine impervious clays and usually underlie the muskegs in the basin. In areas of non-permafrost impervious clays often seal fractures, cracks and shatter zones below depressions, thus preventing groundwater flow and allowing marked muskeg development. By way of reinforcing this point that groundwater flow is negligible in the study region, Brandon (1965) concluded this to be the case from a groundwater hydrology study in the region involving consideration of evidence from physiographic, geological and permafrost conditions, mine reports and chemical analyses of stream water. Conditions at Con Mine, Yellowknife, just south of the basin have been reported by Mr. Beduz, the mine's maintenance superintendent (Personal Communication, April 30, 1973). He states that there is a minimal amount of surface ground water entering the mine and that this is generally because the faults and shears are tight. Underground water flows are occasionally encountered but in the history of the mine these have been quite easily controlled.

Glacial deposits of many types have resulted in the blockage of previously existing drainage channels. This blocking of channels and the glacial erosion of hollows have resulted in the formation of a deranged drainage patterns. Other ramifications of this glacial topography, on water balance studies include the difficulty of determining drainage divides and the inducement of a significant delaying effect between surplus precipitation and stream discharge.

WEATHERING PRODUCTS

Post-Pleistocene weathering of the exposed bedrock areas merits some mention because of its enhancing effect on detention capacities. Post-glacial erosion appears to have been more pronounced on the granites than on the greenstones. Physical weathering affects the granites and the greenstones similarly but chemical weathering is more prevalent on the granites (Landals, 1970, p 21).

Granular disintegration is a very important chemical weathering process affecting exposed granite surfaces. The products of this process are washed into depressions on the bedrock surfaces. This process also leaves the surface of the granite pitted and it is thus able to hold a film of water on the surface thereby increasing evaporation and reducing runoff. It has been estimated that since deglaciation less than one half inch of the granite surface has been removed by granular disintegration as evidenced by comparing granite exposures with adjacent glacially polished and striated dykes (Landals, 1970, p 21). Half an inch is probably greatly in excess of the median amount. The greenstones have not been affected by chemical weathering to an appreciable extent as is indicated by the excellent preservation of abundant striations and chattermarks.

Important physical weathering processes are joint block separation, exfoliation, and especially frost shattering.

SOILS

Soil development or lack of development plays an important role in moisture balance patterns. The thick accumulations of intrazonal organic soils in bogs and swamps are very significant because of their great moisture storage capacity, while the azonal soils developed on bare rock result in much lower storage capacities.

PERMAFROST

As located on MAP 1246A Permafrost in Canada, (Brown,1967,)Baker Creek basin lies within the zone of discontinuous but widespread permafrost. Permafrost is "the thermal condition under which earth materials exist at a temperature below 32°F continuously for a number of years" (Brown, 1969, p 13).

Permafrost has a number of important effects upon water balance patterns. Groundwater that enters a zone of permafrost (which is not already saturated) becomes immobilized as ice and is removed from the dynamic water balance. If the unconsolidated material is already saturated, (as is usually the case in the study area), the permafrost table provides an impermeable barrier to infiltration. This barrier impedes drainage in much the same manner as bedrock and is partially responsible for the extensive tracts of muskeg. Since approximately 60 percent of the study area may be estimated from air photographs to be exposed bedrock and most of the remainder is underlain by permafrost or bedrock at shallow depth, (usually less than three feet), groundwater movement is very limited except on a local scale. The permafrost table also restricts the depth of rooting of trees in many locations.

The active layer strongly influences water balance patterns in the basin. It acts as a reservoir for moisture storage taking up large volumes of water in some seasons and releasing it or holding it for evapotranspiration later. The degree of saturation of the active layer in the spring has a marked influence on snowmelt runoff patterns as this controls the average storage space for moisture uptake. The often saturated and relatively cold condition of the active layer is an important influence in retarding transpiration through vascular plant species.

VEGETATION

Baker Creek basin lies within the subarctic Forest-Tundra Transition vegetation belt. It is on the forested side of the Arctic tree line but the trees are stunted as a result of the harsh climate and there are open moss-lichen areas interspersed with the wooded areas. (Kakela 1969, p 64).

A full description of the vegetation of the southern portion of Baker Creek by plant association has been

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presented by Landals (1970, p 23-42).[*] Two major categories are identified: upland ridge sites and depressional valley sites. Those aspects of the plant associations salient to the present study are summarised here.

UPLAND RIDGE SITES

1. Jack pine-lichen: This association is essentially composed of large expanses of bare rock but for a few jack pine (Pinus banksiana) rooting in the fractures and the small depressions with some material soil. Crustose lichens (spp.) are abundant and widespread as a pioneer species on bare rock.

2. Jack pine-bearberry: The characteristic surficial deposit associated with this association is thin drift which is generally coarse and stony with a low moisture holding capacity. Jack pine is found (25 percent crown cover) and ground cover is generally continuous, being dominated by bearberry (Arctostaphylos spp.) but with lichen locally very abundant. This association is only represented over a relatively small area of the basin and this statement also applies to the aspen-birch and jack-pine associations.

3. White spruce (Picea glauca): Scattered throughout the upland areas are gentle slopes mantled by accumulations of mineral and organic soil. These sites are occupied by

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[*] More information on vegetation cover in the basin will be available as a result of vegetation mapping studies now in progress and to be reported in 1974 by J. Park.

white spruce, (crown cover greater than 75 percent). Alder (Aldus crispa) is the major shrub in this association and the thick ground cover is made up of a variety of mosses and lichens.

4. Black spruce (Picea mariana): Where large depressions have formed in upland areas their floors are often sealed by fine lacustrine clay. This material impedes drainage resulting in a plant association dominated by black spruce. Poor drainage of these depressions also results in the presence of organic soil largely built up by Labrador tea (Ledum palustre) and Sphagnum mosses. Most of the water that collects in these sites during snowmelt is lost in summer through evaporation and transpiration according to Landals. Drying out of these sites in summer lowers the soil moisture content and permits a more mesophytic ground vegetation than is normally associated with black spruce.

DEPRESSIONAL SITES

1. Black spruce - labrador tea: The most heavily vegetated depressional sites are the black spruce muskegs. This plant association, which occupies extensive tracts of muskeg in the basin is greatly dominated by the two species, black spruce and Labrador tea. A thick ground cover of Sphagnum is everywhere present and permafrost is also usually present at a shallow depth in these locations.

2. Willow-ground birch: This second muskeg association is dominated by willow (Salix spp.) and ground birch (Betula

glanduloso). Ground cover consists primarily of *Sphagnum* mosses, sedges (*Carex* spp.) and aquatic grasses. This association is particularly associated with icings* development along the valley adjacent to Baker Creek.

3. Labrador tea: In certain depressional areas Labrador tea is dominant with *Sphagnum* (spp.), cloudberry (*Rubus chamaemorus*) and horsetail (*Equisetum* spp.) locally very abundant.

4. Sedge meadow: Sedge (*Carex* spp.) meadows occupy drainage sites which tend to dry out slightly as summer progresses.

Preliminary taxonomic data from vegetation mapping transects made within a mile of the main study sites are presented in Appendix A.

MAN'S INFLUENCE ON VEGETATION

This survey of the vegetation in the basin would not be complete without some mention of man's influence on the vegetation of the area. This influence is marked in the southern part of the basin especially in the vicinity of the Giant Yellowknife mine. In addition to the damage commonly caused by cutting during prospecting operations, the vegetation adjacent to the mine has also been greatly modified by fire. As a result the organic layer and vegetation cover has been somewhat reduced.

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* The formation of icings has been described by Landals (1970, p.9): "In early winter, stream flow is impeded by freezing in the form of anchor ice at the bottom and floating ice at the surface, while shallow channel sections may freeze solidly. Water still flowing from upstream floods over the frozen surface of the channel as the flow becomes increasingly impeded. Alternate flooding and freezing, continuing most of the winter, may build up layers of ice six feet in thickness." (See also Landals 1970, p 39-42).

Air pollution has also played a significant role in destroying vegetation in the area around the mine. Arsenic compounds were formerly released unchecked into the atmosphere and this has had a devastating effect on much of the vegetation and wildlife in the area.

It is difficult to quantify the influence of such activities on the water balance patterns in the basin. However its existence should be borne in mind when evaluating this study's results and applying them elsewhere.

CLIMATE

Temperature: Latitude and continentality play major roles in controlling temperature. The study area has short summers and long winters. Whereas at the winter solstice Yellowknife receives slightly less than five hours of insolation, at the summer solstice about twenty hours is received. The greater angle of incidence in the summer increases the intensity of solar radiation. There is also less absorption and scattering of radiation by the atmosphere during the height of the summer (Fig 2-1).

The study area is situated approximately 750 miles from the Pacific Ocean, about 650 miles west of Hudson Bay and about 350 miles from the nearest part of the coast of the Arctic Ocean. It is thus far removed from the moderating influence of oceanic water bodies. Also the latter two bodies are frozen for much of the year and thus have primarily continental characteristics..

The annual mean monthly temperature range is thus substantial (almost 80°F). The extreme temperature range in an individual year could reach almost double this level (Figure 2-2). A factor which slightly counteracts the continental temperature regime in the study area, especially in summer, is the proximity of Great Slave Lake. Because of the slower heating rate of the lake relative to the adjacent land in summer and its slower cooling in early winter before the ice cover forms, the land areas adjacent to the lake have slightly moderated temperature ranges by comparison with more removed locations.

From Figure 2-2 it may be observed that significant evaporation and evapotranspiration is on the average confined to the months of May to September inclusive with the major activity taking place in June, July and August. It is probable that the evaporation regime is skewed towards the fall period for the larger lakes owing to the heat storage lag influence.

FIGURE 2-1 ANGLE OF MID-DAY SUN ABOVE HORIZON. YELLOWKNIFE,62° 27′ NORTH LATITUDE.

Source: Kakela, (1969, p 38)

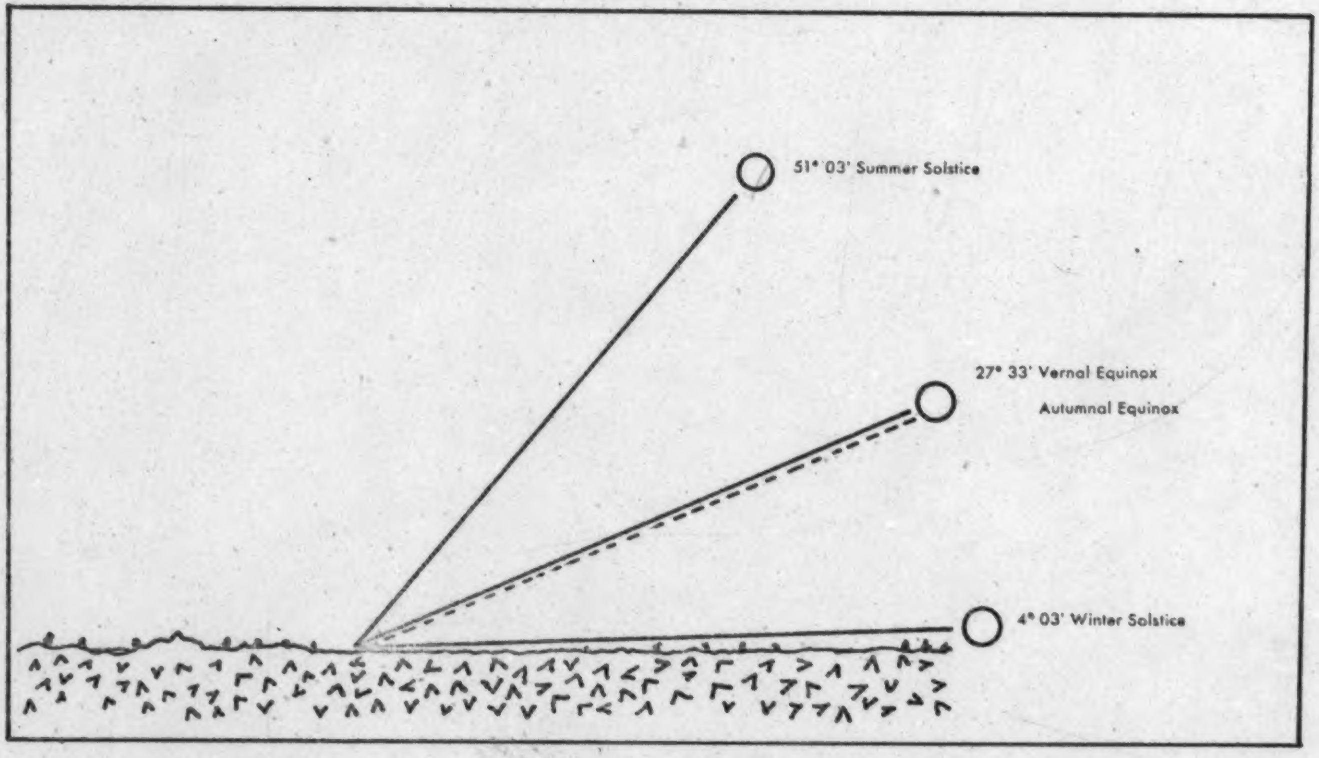
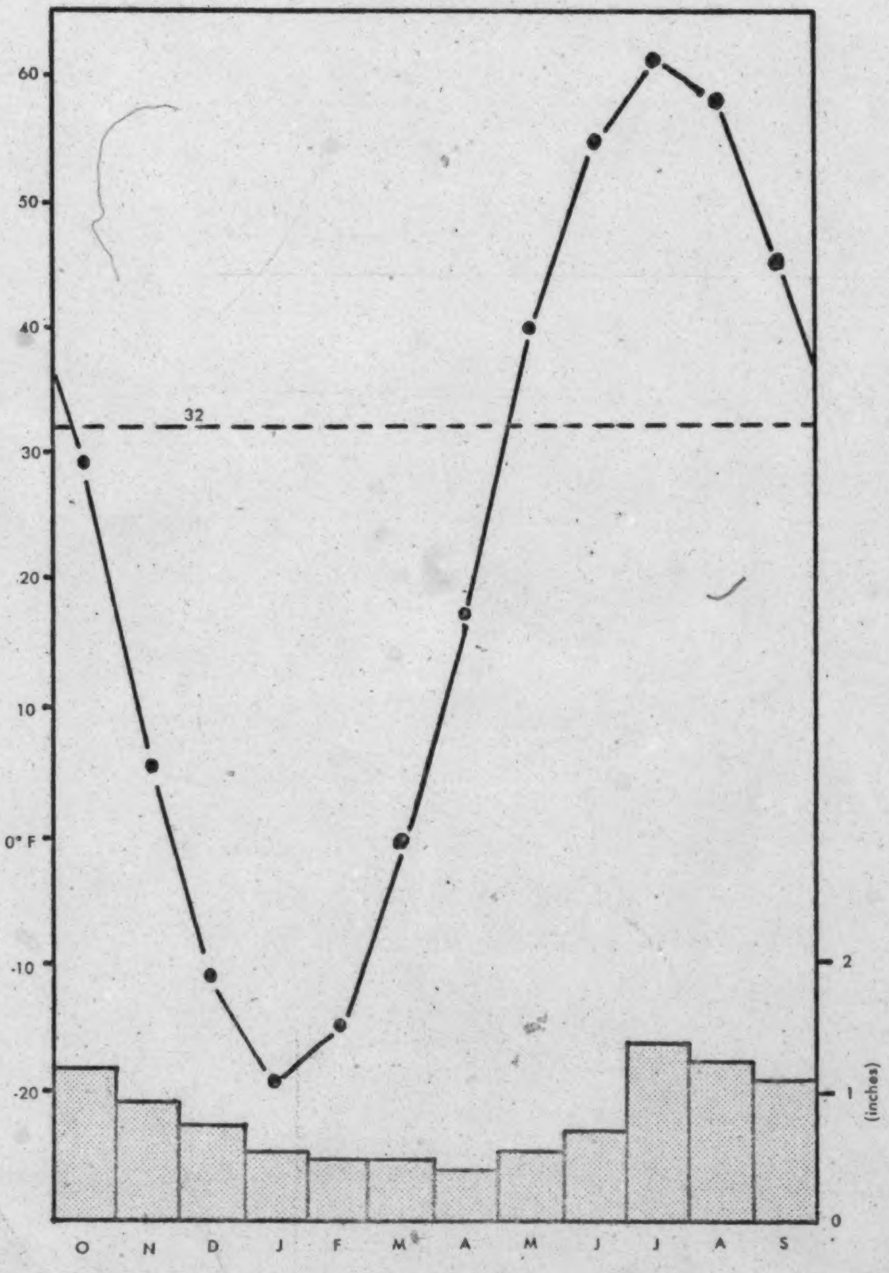


FIGURE 2-2 MEAN MONTHLY TEMPERATURE AND PRECIPITATION. YELLOWKNIFE AIRPORT.

Source: Kakela,(1969,p.40)

MEAN ANNUAL: TEMP.21.9° F; PPCT. 9.47 inches



Yellowknife's altitude is relatively low, (approximately 600' a.s.l.), and this tends to reduce somewhat the diurnal range of temperature, especially in summer, to relatively low levels. Low pressure systems originating mainly in the Gulf of Alaska control the summer weather. Some local low systems tend to develop because of the rapid heating during the short but intensive summer. Also cyclonic storms occur less frequently than in more southerly locations and as a result the sequential temperature fluctuations associated with the passage of cyclones are not as prominent in the study area as they are in mid-latitudes.

Precipitation: In a water balance evaluation the amount of precipitation received provides the input to the system. The study area is located in a region of very low average annual precipitation. Mean annual recorded precipitation at Yellowknife Airport is 9.73 inches (Figure 2-2). This low precipitation total is due to three main factors. Cold air masses are prevalent over the area. They have little moisture holding capacity and thus produce only limited precipitation with uplift. Secondly, there are few precipitation mechanisms to induce release of what little water vapour does occur in the atmosphere. There are no orographic barriers capable of causing significant uplift of air masses. Convectional and cyclonic mechanisms rarely occur in winter and are at best poorly developed in summer. Thirdly, the study area is isolated from sources of marine air. It is situated in a rainshadow location with regard to any westerly flow of air from the Pacific Ocean. However Pacific air masses are still the major source of moisture. Cyclonic and more limited convectional uplift of this air must exceed the levels reached during the mountain passage. This constraint results in the precipitation that falls being moderate in frequency and intensity (Kakela, 1969, p47).

The annual regime of precipitation indicates a moderate concentration in the summer and early autumn. During these periods there is an increased frequency of precipitation mechanisms, in addition to the increased water vapour holding capacity of the warmer air during this period. Conversely however the higher air temperatures correspond to increased demands for moisture by plants and increased rates of evaporation from free water surfaces.

The detailed nature of the precipitation can be very significant in evaporation and evapotranspiration studies, particularly in areas of bedrock or moss-lichen surfaces. For the study area the frequency of precipitation in summer tends to be in the low to moderate category (Table 2-1).

TABLE 2-1

Percentage frequencies of hours with precipitation:
Yellowknife Airport, 1957-1966

	Apr.	May	June	Jul.	Aug.	Sept.	Oct.
Thunderstorms	--	--	0.3	0.6	0.5	--	--
Rain	1.1	4.4	6.6	8.3	8.9	12.1	6.1
Snow	11.0	3.0	0.4	--	--	2.4	19.8

Source D.O.T., Hourly Data Summary No. 78 Yellowknife Airport.

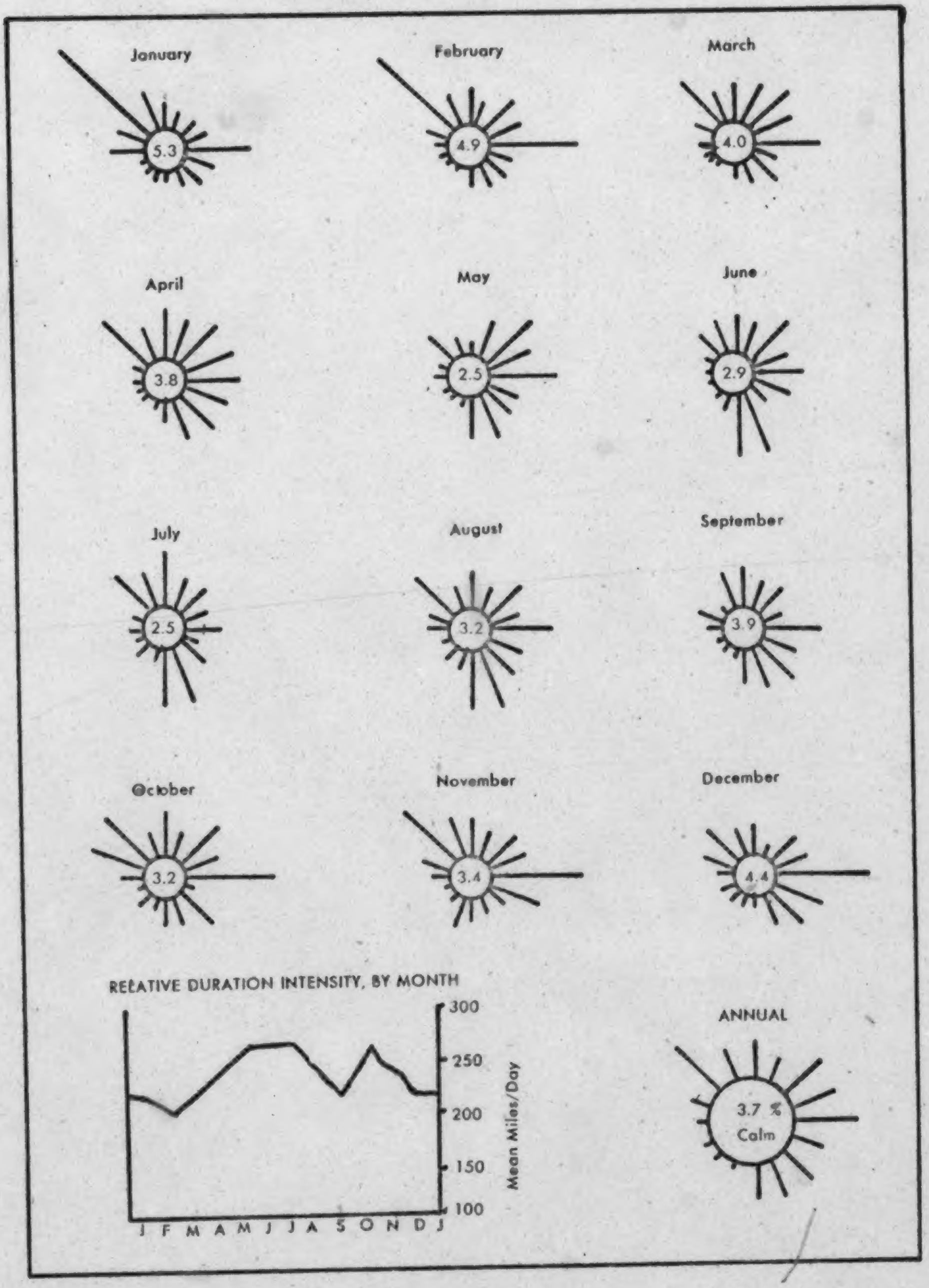
The individual precipitation events are also marked by the paucity of yield and the vast majority of precipitation hours are classed as "light" with very few "moderate" or "heavy" occurrences (Kakela, 1969, p 46). The daily precipitation record for Yellowknife Airport for the months of April to September inclusive from 1966-1972 is presented in appendix form as additional background information on this aspect (Appendix B, Canada D.O.T., A.E.S., Monthly Record 1967-1972).

Winds: Evapotranspiration is influenced to an important degree in the study area by wind velocity and duration.

The twelve monthly and one annual mean duration intensity wind roses from Yellowknife Airport for the 10 year period 1957-1966 are shown in Figure 2-3. In these wind roses the direction of the wind is represented by the orientation of the lines. The hourly duration of wind from anyone direction multiplied by the mean speed of the wind from that direction yields a duration intensity value for the wind blown from that direction in miles per hour. The

FIGURE 2-3. MEAN DURATION-INTENSITY WIND ROSES FOR YELLOWKNIFE AIRPORT 1957-1966.

Source: Kakela,(1969,p.51)



lengths of the lines on the wind roses are proportional to these duration intensity values.

It may be noted that in summer the wind pattern is more variable than in winter with some prevalence from southerly directions. It is probable that the variable pattern is related to the influence of Great Slave Lake in encouraging the development of daily land/water breeze effects.

Climate Classification: Climate classifications for Yellowknife Airport data have been reviewed by Kakela (1969, p 52-55). In most classifications based on that by Koppen Yellowknife is centrally located in the Subarctic class. The point is made that though precipitation is light in Yellowknife's class, evaporation is small, making the climate effectively moist. Recently the Thornthwaite classification system has come to enjoy preference over the Koppen-based systems. (Carter, 1967). This classification is of particular significance here because of its incorporation of the potential evapotranspiration (P.E.) concept. M. Sanderson has applied the Thornthwaite classification to all of Canada (Sanderson, 1948 b). The study area was classified as having a moderately large moisture deficit compared to an existent, but small, annual surplus. The classification also indicated that Yellowknife has a P.E. value, estimated from mean monthly temperature and a length of day factor, which is "low" annually, i.e. between 11.22 and 16.83 inches (Kakela 1969, p55).

The first section of Chapter Five is devoted to a comparison of weather and climate patterns during the study period with longer term trends. Spatial variability of parameters is also discussed.

CHAPTER 3

LITERATURE REVIEW

INTRODUCTION: The literature review is organized into three main sections. The first section is focussed upon evaporation and evapotranspiration from vegetation communities commonly found in the Subarctic. The literature on albedo patterns over subarctic surfaces also receives attention in this section in addition to a consideration of the literature relating permafrost and vegetation interrelationships. Ground heat flux is also considered here. Lake evaporation forms the focus of the second section and basin scale working estimates of evaporation and evapotranspiration are considered in section three.

A reference map indicating the location of various studies mentioned in the literature review has been constructed (Figure 3-1)

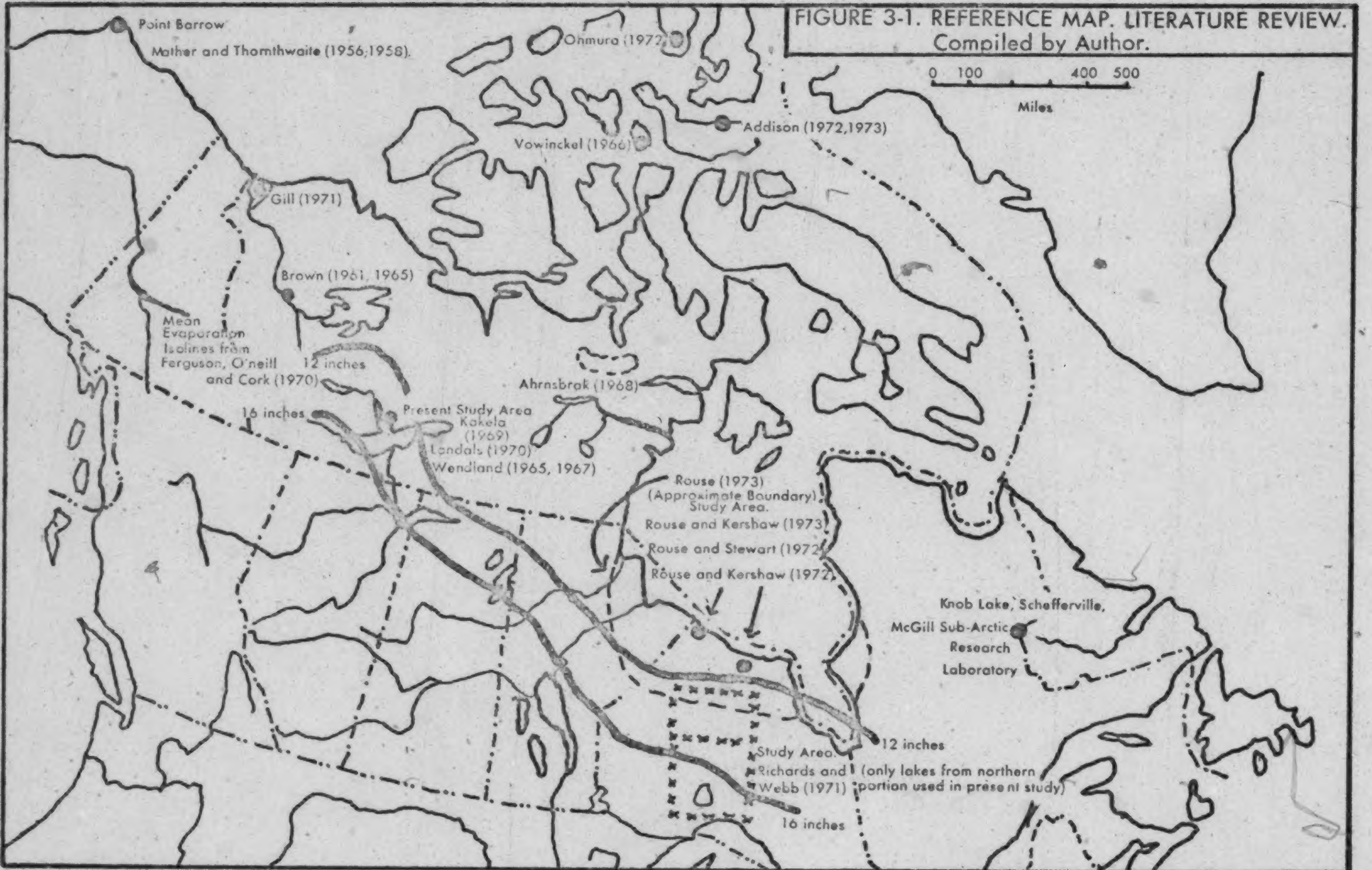

0 100 400 500
Miles
Point Barrow
Mather and Thornthwaite (1956,1958).
Ohmura (1972)
Addison (1972,1973)
Vowinckel (1966)
Gill (1971)
Brown (1961, 1965)
Mean Evaporation Isolines from Ferguson, O'neill and Cork (1970)
12 inches
16 inches
Ahrnsbrak (1968)
Present Study Area
Kakela (1969)
Landals (1970)
Wendland (1965, 1967)
Rouse (1973) (Approximate Boundary) Study Area.
Rouse and Kershaw (1973)
Rouse and Stewart (1972)
Rouse and Kershaw (1972)
Knob Lake, Schefferville, McGill Sub-Arctic Research Laboratory
12 inches
Study Area. Richards and Webb (1971)
(only lakes from northern portion used in present study)
16 inches

FIGURE 3-1. REFERENCE MAP. LITERATURE REVIEW.
Compiled by Author.

EVAPORATION AND EVAPOTRANSPIRATION FROM ARCTIC AND SUBARCTIC VEGETATION COMMUNITIES

In subarctic and arctic regions mosses and lichens comprise a relatively great proportion of the total flora by comparison with mid-latitude areas. These non-vascular plants do not transpire and hence may be expected to significantly influence water balance patterns. The vascular flora which occurs in the north-trees, shrubs, sedges etc.,--cannot be viewed as functioning in a similar manner to vascular plant species found in the south since this flora in the north tends to be associated with wet peatlands and /or permafrost conditions which may be expected to alter plant behaviour.

It is difficult to classify the literature available on this topic. A systematic survey by different plant types is complicated by the comparative nature of many of the studies which have been undertaken. Also, in some cases the vegetation is not indicated, though the results may be of interest. The organisation followed here is a combination of the geographical location of certain studies and systematic consideration of different plant types or different influences on evapotranspiration.

Lichen covers have received a relatively high degree of research attention in Northern Canada by comparison with moss covers. Pioneer work on moisture balance studies over lichen covers was conducted by personnel associated with the McGill Sub-Arctic Research Laboratory, Knob Lake, (Schefferville)on the Quebec-Labrador Boundary in the central portion of the Labrador-Ungava peninsula. The initiative for these early studies may be attributed to Hare (1954, 1955). He noted the evidence of relatively high measured runoff values, (by comparison with measured precipitation), for the subarctic regions of the Labrador-Ungava peninsula and also for similar terrain in Northern Finland. Hare suggested reduced evapotranspiration as a possible explanation pointing to the widespread occurrence of bare rock and the cryptogram floor cover, (mosses and lichens), possessing only rudimentary capacity to draw on soil water. Nebiker and Orvig (1958) attempted a partial proof of Hare's hypothesis. They installed Thornthwaite-style evapotranspirometers beneath a lichen cover in the summer of 1956 choosing open woodland sites at Knob Lake. P.E. rather than actual evapotranspiration (A.E.) was the quantity of water loss measured by these tanks since the soil was kept at field capacity by sprinkling water over the tanks after a 24 hour period without rain had occurred. An attempt was made to reduce the "oasis" effect by also sprinkling the surrounding area with water at the same time.

Measured water loss from the tanks, (i.e. measured P.E.), averaged a total of 5.4 cms. for the months of July and August. The Thornthwaite computed P.E. for the same period was 18.6 cms. Precipitation was 20.1 cms. Nebiker and

Orvig (1958) noted that the level of water loss was "only one-third of the value usually experienced under similar climatic conditions (but with a different vegetation cover)." (p 383) The peculiar characteristics of the lichen cover were cited as the major contributing factor to the very low measured water loss. Lichens and several moss species get their water and their mineral nutrients, with the water, through the cells in the thallus and not through any vascular organs from the soil itself (Hustich, 1951). They die at their lower end. By comparison in all other types of plants used as surface covers in measurements of P.E. the plant systems are complete, i.e. possessing roots, stems and leaves to conduct water into the atmosphere. The lichen possesses none of these parts and are instead like a mat of uncompacted straw with the bottom layer of a more spongy consistency. In the field lichen appears bone dry on the top but soggy and wet just above the humus layer so that moisture is clearly not limiting. As a result Nebiker and Orvig believed that, "there is no 'P.E.' in lichen" (p 383) They further state that the customary methods of computing the soil moisture conditions cannot be used in lichen covered areas. Two main reasons are offered to explain why lichens allow only the very slowest rate of water loss: a) "The reflective power of the lichen cover is exceptionally high, probably near double the 30 percent usually given for an almost dry grass carpet. The heat absorbed by the lichen covered surface is not sufficient to sustain a large water loss". b) "The lichen cover serves as an effective mulch, protecting the soil surface from evaporation." (Nebiker and Orvig, p.383).

Similar studies at Knob Lake during the summers of 1957-59 generally substantiated the 1956 results although measured losses and discrepancies with computed P.E. were not quite so great (Watts, Galloway and Grenier, 1960). In the latter study it was found that measured evapotranspiration was very variable and could be only weakly correlated with climatic parameters. Techniques of watering the tanks and the nature of the lichen surface were found to be of most significance in explaining measured evapotranspiration variability. High rates were observed just after precipitation events.

Hare (1966) has commented on the above results. From personal estimates of net radiation he concluded that only about one-sixth of the available energy went into evaporation according to the 1956 results. Despite the somewhat larger losses observed in the 1957-1959 studies, Hare noted that the measured values still fell short of the net radiation equivalent by a wide margin. From these results he generalised that only about 40 percent of the net radiation was being utilised for evapotranspiration. Findlay (1969) has indicated that average actual evapotranspiration from land surfaces in the Knob Lake area is about 7.5 inches (18 cms). This is approximately 40 percent of the evapotranspiration that maybe expected to occur according to recent estimates of net radiation data for the region (Hare and Hay, 1971, p 91).

Rouse and Kershaw (1971) have recently studied the heat and water regimes of lichen dominated subarctic surfaces in the Hudson Bay Lowlands. Soil moisture measurements made in a mature lichen woodland with thinly spaced trees, and in areas of recent and older burning, indicated that soil moisture under the lichen dominated surface was at least 40 percent greater than in either of the burned areas. This suggested that a mature lichen cover offers a high resistance to the evaporation of soil moisture. Energy budget (Bowen ratio)* determination of the evaporation regime indicated that lichen-dominated surfaces act as an effective mulch in preventing evaporation from the subsurface zone whereas the burned areas which are able to evaporate more water into the atmosphere when moist, also develop strong resistance to evaporation as the soil surface layers become drier.

The workers found that evaporation reached a peak during the high sun period when the direct solar beam is best able to penetrate the lichen mat and tap the soil moisture reserves. The average Bowen ratio for all the measurements from the lichen surface was 0.61, only 62 percent of the available energy being used for evaporation. Another experiment at this site indicated that the lichen canopy, after being saturated from a heavy rain shower, exhibited a considerable resistance to evaporation being particularly strong in the morning but noteworthy throughout the day. Canopy evaporation was observed to progress by layers from the top down. At the old burn site the average Bowen ratio was 0.71. Soil moisture measurements at this site averaged 68 percent of field capacity during the observation period whilst the soil just below the lichen mat was at 99 percent of field capacity.

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* See Appendix C

A major implication of the study is that there will be substantially less summer runoff, both surface and subsurface, from the burned areas compared with the lichen-covered surfaces. It is not known if this implication would extend to burned cover moss-dominated surfaces such as those which occur in the present study area. It seems probable however that burning would remove most of the living moss cover and reduce the efficiency of the mulching effect. The greater exposure of the moist soil layers will favour increased evaporation with a concomitant increase in storage capacity to be filled, at the onset of precipitation, before runoff may occur.

Rouse and Kershaw (1973) have made other studies of the relationship between soil moisture and evapotranspiration at lichen dominated beach ridge sites on the Hudson Bay coastline adjacent to East Pen Island. Evapotranspiration was estimated by the soil water budget method. Selected results are reproduced in Table 3-1. According to Rouse and Kershaw the results were that the wettest surfaces and

TABLE 3-1

SEASONAL EVAPOTRANSPIRATION: HUDSON BAY COASTLINE (NEAR E. PEN ISLAND)

JULY 7 - AUGUST 17 - 1971

SITE	Percent vol. Soil Moisture (cms.)	Change in Soil Moisture	Precipitation (cms.)	Evapotranspiration (cms.)
1. Young Beach Ridge 49 percent ground cover of *Dryas integrifolia*	8	-0.10	8.71	8.61
2. Older Beach Ridge 66 percent ground cover of General lichen heath association	25	+5.00	9.44	13.44
3. Blowout on sandy ridge crest. Almost totally unvegetated	22	+6.60	8.44	15.04
4. Poorly drained lowland. 82 percent ground cover. 50:50 Vascular:lichen Permafrost at 90 cms.	38	+1.20	8.44	9.64
5. Oldest Beach ridge Approx. 95 percent cover of lichen. Permafrost at 40-60 cm.	42	+0.40	8.28	8.68

Source: Rouse & Kershaw (1973).

the very driest surface yield the least moisture. In the very wet areas the soils remain cold during the summer and much of the absorbed radiant energy is used to melt the active layer above the permafrost leaving only small amounts for the process of evaporation. It may be concluded that the thick non-transpiring lichen at Site 5 presents a physical barrier to a water vapour flux across the atmospheric boundary layer. The dry soils at Site 1 were indicated to have a very low capillary conductivity which is partly a function of volumetric moisture so that in the absence of transpiring vegetation on the bare areas and the dominance of highly zerophytic plants in the vegetated patches the available water experiences difficulty in moving upward through the soil to the earth-air interface. It was stressed that the rigour of the evapotranspiration in the study environment was purely relative in comparison to moist vegetative surfaces in temperate latitudes so that even the most freely evaporating surface still presents a considerable resistance to the upward movement of water vapour. It was also indicated that the high evapotranspiration at the blow-out site may well have been due to its serving as a natural amphitheatre which receives an excess of solar radiation. A more probable explanation is that the evapotranspiration value is erroneous. The assumption was made that net vertical soil moisture flux across the terminal depth of measurement at all sites was zero. However the authors state that this assumption is only definitely valid at the sites underlain by permafrost (Sites 4 and 5, Table 3-1). They admit to reservations about the evapotranspiration estimates for the sites without permafrost. It would appear that the reservations should have been re-iterated and acted upon in their discussion of the seasonal evapotranspiration estimates since this questionable assumption probably explains the very high levels at the blowout site and also for site 2. It is more likely that the large changes in soil moisture were caused by subsurface runoff from the sandy sites rather than evapotranspiration. Site 1 is an anomaly in that it is not underlain by permafrost but it experienced a slight increase in volumetric moisture content. However the most likely explanation for this increase is net subsurface inflow from the higher sites, such as sites 1 and 2. It is therefore probable that this site experienced less evapotranspiration than is indicated.

The application of the above research to the present study area is limited by the relatively low volumetric soil moisture contents at the sites. This is in contrast to the very moist, dominantly organic, soils with 100 percent ground vegetation cover investigated in the present study. Also the level of summer precipitation in the present study area is significantly less than that for the Hudson Bay Lowland. As will be shown later the level of summer precipitation tends to be closely correlated with the level of evapotranspiration from sites with extensive hygroscopic vegetation cover and where drainage is poor.

Ahrnsbrak (1968) has conducted a study of the summertime radiation balance and energy budget in the central part of the Keewatin, N.W.T. (Fig. 3-1). All his sites were vegetated with a lichen mat several centimetres thick. The soil at each site was largely sand although one site was more gravelly and also very wet. The results were that during the summer net radiation is nearly balanced by sensible heat transfer. During July Ahrnsbrak estimated that the soil heat flux into the soil accounted for approximately 15 percent of the energy budget but that by August it becomes negligible. During the entire season the transfer of latent heat over the surfaces studied was seen to be very small.

Ahrnsbrak acknowledged that the very small values for the transfer of latent heat to the atmosphere in his study area, one of much open water, seemed paradoxical. However he demonstrated that crude water balance considerations also indicated this component would be very small in value. He noted the annual precipitation for the area to be less than 20 cms. and assumed no runoff to occur. Therefore annual evaporation could be taken as 20 cms. and, "if all this were to take place during the three summer months in the region this would imply slightly over one* mm of evaporation per day, or 60 langleys per day, or less than 20 percent of the net radiation. This however is an average value for all surfaces. In actuality most of the evaporation is probably taking place over the lakes and other standing water, and evaporation over the relatively drier land surfaces is most likely much less " (Ahrnsbrak, 1968,p15).

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* It would appear that Ahrnsbrak's figure for evaporation per day should be two mm instead of one mm. The effect of this error is probably counteracted however by rejecting the assumption that no runoff occurs. Approximately ten cms. of runoff may be estimated for this region from Figure 2, p 84, in Hare and Hay (1970).

Evaporation through moss covers in arctic & subarctic regions has not been studied very widely. Ground energy exchange studies at permafrost sites at Norman Wells N.W.T., conducted by Brown (1961, 1966), produced some initial information on this parameter. His study is particularly important for illustrating differences in evapotranspiration between a temperate latitude grass species introduced at Norman Wells and indigenous vegetation such as moss, lichen and sedge.

Sanderson (1950) installed Thornthwaite type evapotranspirometers at Norman Wells partly to test validity of Thornthwaite's empirical method for computing P.E. for high latitude sites. A critical factor however was that the tanks were sown with an exotic vegetation species, Kentucky Bluegrass. This factor is mainly responsible for the satisfactory performance of the Thornthwaite formula in computing a P.E. value which closely corresponded with that measured at the evapotranspirometers.

During the summers of 1959 and 1960 Brown (1961,1965) conducted further P.E. investigations at the Norman Wells evapotranspirometers site. In addition however he also investigated P.E. at four other sites which included the various types of vegetation growing naturally in the region: sedge; sparsely forested moss cover; open moss cover, and open lichen cover. At the latter sites Brown used less sophisticated oil drum and juice can evapotranspirometers. These necessitated periodic sprinkling of water in excess of the anticipated P.E. onto their surfaces. It is not clear whether an attempt was made to obviate the oasis effect at these sites by also sprinkling water on areas surrounding the tanks.

The main period of investigation extended from June 30 to September 17, 1960. (80 days inclusive). The oil drum instruments were considered to probably give the more realistic P.E. values by comparison with the smaller containers because of their larger surface area, greater depth and decreased rim effect. Results from the study are reproduced in Table 3-2.

From the results Brown concluded that P.E. rates were significantly higher through the Kentucky Bluegrass at the Thornthwaite site compared with the P.E. rate at the other sites. The disregard in using the Thornthwaite formula of different plant types seemed to Brown to be an objection particularly for northern areas where moss and lichens grow widely. He noted that Sphagnum and other mosses are strongly hygroscopic and possess an ability to lose moisture rapidly and in large quantities when wet. Elsewhere, Brown (1966, p21)

TABLE 3-2

EVAPORATION & EVAPOTRANSPIRATION

AT NORMAN WELLS, N.W.T.

JUNE 30 - SEPT. 17, 1960

Site:	1	2	3	4	5	D.O.T.
Vegetation:	K. Bluegrass	Sedge	Moss (trees)	Moss (no trees)	Lichen (no trees)	Airport met. station
Rainfall: (cms.)	17.89	18.63	18.89	19.30	19.10	19.06
P.E. (cms.) oildrums	22.90	12.51	11.76	14.70 (average of three)	10.75	--

Thornthwaite tank	19.41 cms. (Pan B)
Class A Pan	25.63 cms.
Atmometer	19.47 cms. (from the formula by Robertson LEX0.0034 x 2.54 = P.E. (cms.))
Thornthwaite formula	24.75 cms.
Penman formula	19.32 cms.
15 cms. diam.pan.	20.32 cms.

Source: Brown (1961 p 126).

has written: "Sphagnum especially, and to a lesser degree true mosses, tend to absorb and raise water in the manner of a lamp wick". Regarding the lichen, Brown observed that it was not certain whether the transfer of water vapour from the wet basal layer to the atmosphere above the lichen takes place by evaporation or merely by diffusion exchange. The area of vegetative surface per unit ground area of moss and lichens was also mentioned by Brown as being possibly very different from that for higher plants (1965,p20).

Brown strongly questioned the reliability of sprinkling water on the tanks in maintaining moisture conditions at field capacity. He noted that this practice changed the climate of the vegetation under test and altered the natural moisture regime. In addition the use of very small test plots such as oil drums for measuring P.E. because the edge effect is so pronounced that it probably changes the evapotranspiration. In view of these questions it may be appreciated that actual evapotranspiration from the sites would probably have been significantly lower than the values presented in Table 3-2.

The precipitation values in Table 3-2 appear to be significantly higher than those normally encountered for the same period in the present study area. It is surprising that greater evapotranspiration was not associated with this high level of precipitation. It must be assumed that the bulk of the precipitation fell as very heavy rainstorms, much in excess of the storage capacity of the top moss cover and with heavy infiltration. An associated possibility is that the evapotranspirometers provided excessively well-drained conditions. Under natural conditions it is probable that the bulk of the precipitation would be held in the moss or top soil layers for evaporation when drying conditions returned. Conclusions derived from Brown's results must therefore be attached to reservations about their validity.

Several instances may be noted in the literature where attempts have been made to relate the occurrence of a high permafrost table to high evaporation or evapotranspiration rates through the moss or lichen covers. Tyrtikov (1959) has reviewed the position of "the majority of geocryologists," who, "following M.I. Sumgin, postulated that moss is a strong cooling agent of the soil, mainly because of the fact that it evaporates a large quantity of water and therefore strongly minimises the amount of heat which can go to warm the soil " (Tyrtikov, 1959, p5). Tyrtikov stated that not one single experiment had been furnished to confirm this suggestion. He pointed to apparent limitations to evaporation from sphagnum moss covers suggested by Myshkovskaya (in Tyrtikov, 1959). Her observations showed that the most hygroscopic sphagnum mosses were not only capable of evaporating moisture but also of maintaining moisture at other times in opposition to conditions favouring evaporation. The low thermal conductivity of the dry layer of the moss apparently has an important effect in retarding warming of the soil.

Brown (1966, p 21) has also mentioned the possibility that the high permafrost table under true moss, sphagnum and lichen may be caused in part by high evaporation rates that prevent large quantities of heat from entering the ground. However his observations at Norman Wells showed that rates of water loss from moss, lichen and sedge were lower than that through the bluegrass site discounting the likelihood that high permafrost tables were associated with high evaporation rates. Pihlainen (1963, p 26) has also suggested that moss and lichen covers may be able to protect the underlying soil against heat gain, "more by an insulating or albedo effect, rather than by an evaporative cooling effect."

Myshkovskaya (1913, in Tyrtikov, 1959) obtained evaporation values for different species of moss and also lichen. No mention is made of the method used to obtain the results which are reproduced in Table 3-3.

TABLE 3-3

APPROXIMATE AVERAGE EVAPORATION OVER THREE SUMMER MONTHS; BOMNAK, USSR.

	Forest				
Vegetation cover	Poly-trichum	Sphagnum	Sphag. and Poly-trichum	Poly-trichum	Lichen
Evaporation rate (cms./month)	5	10	8	10	55

Source: Myshkovskaya, 1913 (in Tyrtikov, 1959, P 27).

The table indicates that various vegetation associations evaporate varying quantities of water and that even the same cover evaporates about twice as much in an open area as in a forested area. In view of the doubt over the methods employed it is perhaps best to interpret Table 3-3 as indicating relative variations rather than absolute levels of evaporation. Bomnak Station (55°N, 130°E) is located in Eastern Siberia about 250 miles inland from the Sea of Okhotsk. It lies within the discontinuous permafrost zone and its climate is classified as D.w.c. on the Köppen-Geiger System (Strahler, 1969, Plate 2). It has thus a rather similar summer climate to that experienced in the present study area except that it is perhaps slightly wetter.

A Russian group has more recently studied surface and internal evaporation from and in 'highmoor bogs' at Lammin-Suo Station* in the Vyborg district of Leningrad Province of the U.S.S.R., close to the most southern part of the border with Finland (Belotserkovskaya et al, 1969). The methods involved detailed heat and moisture balance monitoring. The microsites studied were a pool and a string bog ridge which rose on average about 26 cms. above the pool.

A dense cover of moss (various Sphagnum species) vegetated the ridges in addition to some well-developed shrubs, and a few small pine trees. Cotton grass and cloudberry predominated in the grass story. In the pools a moss grass cover was present. The moisture content by weight in the top 35 cms. of the peat at both sites averaged about 90-95 percent. The ground water table during the summer averaged 40-45 cm. below the surface on the ridge and 20 cm. at the pool. The results are reproduced in Table 3-4.

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* Lammin-Suo bog station lies at about 61°N, 28°E. Its climate is classified as D.j.b. on the Köppen-Geiger System and is thus roughly similar to that for Northern Alberta. During the growing season the normal evaporation for high moor bogs in this region is about 30 cms (Bavina, 1967, p 369).

By way of interpretation the group noted that during the relatively wet conditions in July 1963 evaporation from the pool areas was significantly higher than that from the ridges. During the dry July of 1964 there was little difference between the sites in terms of evaporation, and in one decade evaporation from the ridge was greater than from the pool. This latter pattern was explained by reference to the ground water levels in the pools dropping to below 25 cm. At this point the moisture content by volume of the moss is only 23 percent and does not exceed the intracellular moisture capacity. Under these conditions the light coloured mosses of the pools have a much higher albedo (of about 27 percent) than the relatively dark vegetation of the ridges where the albedo is 17-19 percent (p 557).

The group also monitored internal evaporation and condensation patterns between three surface layers at the sites. It was found that for the observed ground water levels that from 80-100 percent of total evaporation could be ascribed primarily to the transpiration of the bog

TABLE 3-4

10 DAY AVERAGE EVAPORATION FROM LAMMIN-SUO STRING BOG STATION

VYBORG DISTRICT, U.S.S.R.

JULY 1963

Decade	Evaporation (mm./day)		Ground water level (cms.)		Evaporation (mm./day)		Ground water level (cms.)	
	Ridge	Pool	Ridge	Pool	Ridge	Pool	Ridge	Pool
I	3.0	4.0	-38	-13	1.8	2.1	-58	-33
II	2.4	3.0	-51	-26	2.1	1.9	-60	-35
III	2.9	3.5	-44	-19	2.5	2.8	-62	-37
Approximate monthly total cms. (inches)	8.3 (3.3)	10.5 (4.2)	-- --	-- --	6.4 (2.6)	6.8 (2.7)	-- --	-- --

Source: Belotserkovskaya, Largin and Romanov; 1969, p 554.

vascular plants and that net internal evaporation from the surface played an insignificant role in the total evaporation process. Since the precise methods involved in this research are not detailed in the paper it is probably wise to view the results with some skepticism. As later interpretation will indicate however a good modicum of consistency may be attributed to the results.

The vascular flora present in arctic and subarctic regions experiences growth restrictions which can often be traced to permafrost conditions. Tyrtikov (1959, p 15) has summarised the two main influences of permafrost on vegetation as: a) Impeding the warming of the soil during the growing period to the extent that the temperature of the root zone is considerably below optimal. b) "Swampifying" of the soil leading to a decline in aeration and an impoverishment of nutritive substances. (Tyrtikov is apparently generalizing only for high moisture content sites, (eg. bogs), affected by permafrost and not for drier sites affected by permafrost).

The low temperature of the soils over permafrost had been thought to cause physiological drought, i.e. retardation of the absorption of water by the roots so much that they cannot satisfy the requirement of the above ground organs of the plants which leads to their destruction. Tyrtikov points out however that certain experimental investigations are not in agreement with this theory (Dadykin 1954).

These investigations have associated the low temperatures with a retardation of the rate of biochemical reactions and in particular of reactions of the synthesis of albumens and amino acids which occurs in the roots and without which living cells can not be built.

Tyrtikov (p 16) notes that the aeration and nutritive deficiencies associated with soils over permafrost also weaken the development of plant root systems, but he adds that this whole aspect has been barely studied. The operation of the transpiration process through plants growing in conditions influenced by the presence of permafrost does not appear to be well understood other than general statements to the effect that permafrost weakens the development of life activity of the subsurface and above ground organs of the plants (Tyrtikov, 1959, p 23). A recent review by Bliss (1971) does not record any significant advances in this area up to 1971.

Trees and shrubs constitute an important part of the vegetation communities in the present study area. It was not possible to undertake field research to establish transpiration rates through these plants. The following brief literature review indicates possible patterns.

Hare (1966, p 10) postulated that the scattered trees and shrubs in the woodland zone of Labrador-Ungava suffer excessive transpiration due to their exposed position enabling them to acquire extra energy by advection. He has noted abundant signs of dessication in the trees and shrubs, even in moist areas. No measurements exist to verify Hare's contention. The association of abundant dessication evidence and moist conditions could be used as an argument for restricted, rather than excessive, transpiration through the trees. See later references to Bliss (1960) and Warren Wilson (1957, 1959).

Fraser (1956) has pointed to several influences attributable to the lichen covers that may be limiting rather than encouraging transpiration from trees and shrubs. Almost all the tree roots were observed by Fraser to be confined to the basal layers of lichen or the underlying peat which in turn was observed to remain at or near field capacity for the duration of the summer season. Low temperatures and very poor aeration characterised these soil layers indicating, according to Fraser, the possibility of physiological drought. Fraser also suggested that acids produced by the lichens, and which are nearly insoluble in water, could possibly be toxic to other plants. This factor could also contribute to a reduction of the rate of absorption of soil moisture by the tree roots.

Bliss (1960) has reported a study on the transpiration rates of arctic shrubs which appears to be very relevant to the present discussion. It was noted that although the shrubs are somewhat protected their relatively large transpiration surface is in contrast to the much smaller leaf surface of the low grasses, sedges, forbs and dwarf heath shrubs which dominate tundra regions (p 386) Rates were determined by the potometer method. Of particular interest here is transpiration through the arctic (rather than alpine) shrubs from Umiat, Alaska. The data obtained indicated that on a diurnalbasis there is a direct relationship with cloud cover and wind, when speeds are about 6 to 8 mph. At wind speeds below 6 mph. the relationship with water loss was found to be direct. At the higher wind speeds it was concluded that the stomates undoubtedly closed, effectively reducing transpiration. These observations on wind effects support work by Warren Wilson (1957, 1959). He found that while evaporation rates around tundra plants were closely related to windspeed he believed that wind is more important in lowering leaf temperatures than in promoting excessive transpiration with a resultant water balance deficit within the plant.

Bliss (1960) noted that the rates of water loss measured in his study seemed very low when compared with rates for temperate plants. Meyer and Anderson (1952) cite figures of 0.5 to 2.5 cc/dm^2/hr. for broad leaved plants of temperate regions under conditions favourable to stomatal transpiration The figures from Bliss's study ranged from 0.006 to 0.048 cc/dm^2/hr. It is quite probable that such discrepancies are maintained for subarctic trees and shrubs compared with temperate species, though perhaps the differences may not be quite so pronounced. Smaller vascular plants may be expected to behave similarly since it was apparently Bliss's hypothesis that the greater leaf area of shrubs would favour significant transpiration rates in contrast to the smaller plants.

Mather and Thornthwaite (1956, 1958) have reported on heat and moisture balance studies at Point Barrow, Alaska from 1956-1958. Measurements were made mainly at a level, lowlying area with low centered polygons. The only mention of vegetation characteristics at the site was that is was 'grassy', (1956, p9), suggesting vascular rather than cryptogramic flora. It was concluded from the heat balance studies that:

> "The evaporative flux is about 40 percent of the net radiation. In mid latitude areas with moist soils approximately 80-90 percent of the net radiation is used in evaporation of water. It is felt that conditions approximating P.E. existed most of the time at Point Barrow yet only about 40 percent of the net radiation was utilized in the evaporation of water. We assume that the lower evaporation is related to the occurrence of permafrost. The reason for the different rate of utilization between mid-latitudes and the Arctic region needs to be investigated more fully " (Mather and Thornthwaite, 1958,p88).

Juice can evapotranspirometer measurements were in 'fair agreement' with computed evaporation flux for many individual daily comparisons (p80-81, 1958). Moisture levels in these cans were kept at field capacity by precipitation; or irrigation if drying conditions prevailed. In an experiment at another site (high centered frost polygons) two separate groups of small juice can evapotranspirometers were operated, one group being watered periodically to ensure approximation of P.E. conditions, while the other group received no water except from precipitation during the summer season. There was only a small difference (approximately 12 percent) less from the unirrigated group between the measured evapotranspiration from the two groups. This was explained by reference to the fact that the precipitation was fairly well distributed throughout the season entailing only a slight difference in the moisture content between the two groups (1958, p 91).

There were indications that there should be no great drying of the surface soil during the season and this was borne out by observations of generally moist conditions existing in the area during the summers. In addition it was noted that, in all three years of the study period, there was a virtual balance between the precipitation and the evapotranspiration during the summer period (p 93).

The Devon Island, N.W.T., I.B.P. High Arctic Ecosystem project has resulted in some detailed heat and water balance studies pointing out microsite variations (Courtin 1972, Addison 1972; 1973). Some conclusions from research at a sedge meadow merit particular mention in the context of the present thesis.

It was found that the soil at the meadow community was saturated to the surface, and water lies in small shallow pools for most of the growing season. In attempting to explain the relatively low evaporation rate from the Carex in such moist conditions Addison mentioned the shallow gradient of water vapour density between the air and leaves caused by the high relative humidity and low leaf temperature, (1973, p 47). He also noted that part of the explanation appeared to lie in the reduction of root permeability as a result of cold and poorly aerated soil. This would cause a limitation in the flow of water to the leaves, and contribute to low water potentials and stomatal closure (1973, p 82).

For the moss microsite at the meadow Addison noted that since the bases of the moss stems are in water, water movement to aerial parts is possible by capillary action on the outside of the stems (1973, p 101). This process was held to account for the almost 50 percent of net radiation going into evaporative flux at this site. The free standing water site was noted as being the only site where latent heat flux was the major component of net radiation. The observations at the meadow community generally indicated very inefficient operation of the evaporation process.

Other arctic studies which may help to establish trends are those by Vowinckel (1966) and Ohmura (1972).

Vowinckel (1966) has shown that the July energy budget at Resolute Bay N.W.T. is one in which the net radiation is nearly balanced by the storage term, with the transfer of sensible and latent heat accounting for only 10 percent of the energy budget. By August however the sign of the storage term reverses and heat exchange with the atmosphere is balanced by net radiation and heat from the soil. The transfer of latent heat to the atmosphere appears to be very small for Resolute. No mention is made of the measurement site's characteristics.

Ohmura (1972) has reported on another heat and water balance study on arctic tundra on Axel Heiberg Island. Again no mention is made of the surface characteristics of the measurement sites though it may be inferred that at least some vascular flora was present since the term 'evapotranspiration' is employed. The study involved heat balance measurements which allowed solution of the heat balance (Bowen ratio) equation for evapotranspiration, and also lysimeters to measure this component more directly. The latent heat fluxes obtained by both methods were"in very good agreement". (p 175).

For five two-week periods from June-Aug. 1970 the measured heat balance indicated that the evapotranspiration component comprised on average 27 percent of the net radiation, the range being from 12-39 percent. The Bowen ratio varied from 1.2 to 6.0 for the five study periods and averaged 2.5 over the whole period. The evapotranspiration measured at the lysimeters for the period July 5 - August 23rd totalled 4.3 cms. Precipitation was 7.6 cms.

Rouse (1973) has published data on the daytime energy balance of three surface types characteristic of the Hudson Bay Lowlands a clear day in July 1972. His data are reproduced in Table 3-5.

TABLE 3-5

DAYTIME ENERGY BALANCE TOTALS, JULY 4, 1972, HUDSON BAY LOWLANDS. (cal. cm^2)

	Net Radiation	Ground Flux	Latent Heat Flux	Sensible Heat Flux	Bowen Ratio
Upland Ridge	410.4	33.8	164.6	212.0	1.29
Lowland Swamp	488.7	40.1	324.9	123.7	0.38
Lake	542.7	267.9	198.6	76.2	0.38

Source: Rouse (1973)

For the land surfaces the lowland wet site (muskeg/bog) evaporated twice as much water as the upland ridge (lichen-dominated). Evaporation is not particularly efficient for any of these surfaces, including the lake, in terms of the available radiant energy. Albedo is partly responsible for the differences in evaporation from these surfaces and this factor will now be considered in more detail.

ALBEDO:

Pioneer work into the determination of albedo patterns over subarctic surfaces has been carried out by Davies (1962). Airborne albedo measurements were made during the summer of 1961 in Central Labrador-Ungava. Representative values of albedo for the main summer cover types are presented in Table 3-6. Davies noted no considerable variance of albedo between cover types attributing this partly to the generalised albedo sensed by the instruments used and partly to the prevailing drabness of the landscape which is brightened only by lichen in open woodland and tundra expanses. (p 39).

Davies attributed the relatively high albedo of the tundra to its smooth surface, dryness and unconcealed growth of lichen. The relatively high albedo of the burns seems a little surprising since lichen is absent and the surface is notably rough and grey in overall appearance. It was pointed out however that the spruce trees, stripped of their bark, reflect strongly and are particularly noticeable under direct sunlight. Also, many dead trees litter the ground providing good reflecting surfaces for radiation incident upon them.

Open woodland, including extensive lichen surfaces, had been ascribed fairly high albedos in the literature prior to Davies' findings and those of Jackson(1959). (See review of Nebiker and Orvig, 1958). Measurements indicated that the brightness of the cover type had been

TABLE 3-6

SUMMER ALBEDO BY COVER TYPE
FOR CENTRAL LABRADOR - UNGAVA

	Percentage				
Cover type:	Mean	Mediam	Mode	Range	Representative value
Tunda	14.9	14.9	14.4	9-21.5	16.9
Burn	12.4	12.0	11.7	7-17.5	12.4
Open Woodland	12.2	11.5	10.5	6.5-18.5	7-20
Bog & Muskeg	11.0	10.5	9.7	5-17	9.7
Closed Crown forest	9.6	8.9	8.5	4-17	9.2
Water	7.6	7.0	4.8	2-24	--

Source: Davies, 1962, p 83 and p 37.

overemphasised. Ground measurements over the lichen cover did however produce relatively high albedos in the range 15-20 percent (p45). The great range tabled for this category is due to the presence of areas of bare rock, varying densities of ground shrubs and a wide range of tree spacing. (7 percent albedo refers to a vegetation condition where all lichen is obscured by a shrub layer, and 20 percent refers to the sparsest open woodland).

The albedo measurements for bog and muskeg indicated fairly high values bearing in mind the saturated condition of this type. However Davies noted that water was seldom present above the vegetation surface and was thus not visible from the air. He affirmed that the bog and muskeg were essentially vegetation surfaces.

Ground measurements of albedo were undertaken over a water-logged sedge meadow and a value of 16.2 percent was obtained. Davies (1962, p77) took this to support his hypothesis that although a relatively continuous water layer covered the peat and grass root surface layer of a bog, the strong grass growth shaded it, thus the effect of water content on albedo was not very large.

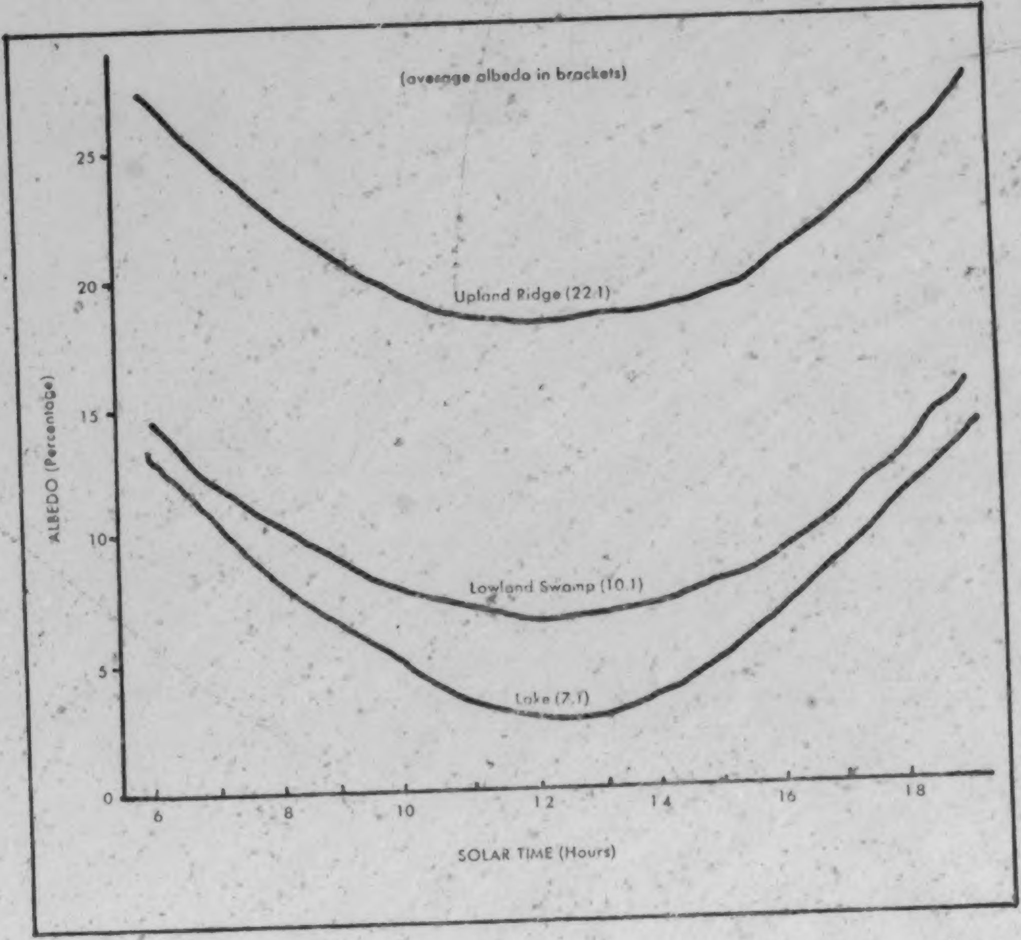
Rouse (1973) has published a diagram relating albedo and solar time for the surfaces covered in Table 3-5. (Figure 3-2).

Wet swampy areas can be seen to take on more the albedo characteristics of open water than of the drier land surfaces. The pronounced variation of albedo with solar

FIGURE 3-2. ALBEDO OF DIFFERENT TERRAIN TYPES.

Hudson Bay Coastline. (Lat.56° N)

Source: Rouse (1973)



time is to be noted for the lake surfaces. This variation renders a mean value for this surface as somewhat meaningless. It is probable that in more northerly latitudes the curves will take on a more accentuated pattern.

The most direct expression of albedo differences lies in net radiation values. Rouse (1973) found that net radiation varied in the proportions 100:90:76 from the lake to the swamp to the ridge.

In the course of Brown's study some crude spot measurements of net radiation were made during August and September 1959 (Brown 1965,p 22). A general qualitative pattern was evident. Net radiation values for lichen were the lowest on clear days but not on partly cloudy or overcast days when differences in net radiation from one plant type to another were small. It was noted however that net radiation values for sedge which is also a light coloured plant were much higher than for the lichen, (disputing Davies' findings somewhat), on clear days and higher than the darker coloured plants such as grass and moss. Net radiation through Sphagnum and other mosses was highest on clear days and on certain overcast days.

Brown noted that despite the apparent differences in net radiation through moss and lichen these plants maintained the permafrost table at about the same level in a given area and near surface ground temperatures were similar under both plant types: "Consequently, if lichen rejects a higher proportion of the net radiative flux than moss this may be compensated by the moss rejecting a higher proportion than the lichen of some other component of the energy exchange " (Brown 1965, p 22).

GROUND HEAT FLUX

It is a common practice in most energy budget studies to assume that ground heat flux accounts for an almost negligible proportion of the net radiation flux by comparison with the sensible and evaporative heat fluxes. However, the detailed mechanism of heat and mass transport in organic soils overlying mineral soils and/or within the active layer of permafrost is poorly understood (Nakano and Brown, 1972, p 20). Therefore it is possible that the assumption of negligible ground heat flux, although apparently valid for the relatively dry mineral soils typical of mid-latitudes, may be invalid for the extensive peatlands which, as Brown and Williams (1972, p 3) have pointed out, attain their optimum development in the discontinuous permafrost zone.

It was not possible in the present study to obtain accurate values of the ground heat flux owing to the lack of the necessary sophisticated flux metering instrumentation. Some limited information on the parameter may be deduced from studies of active layer development at permafrost sites and this is reported later; however the major source of ground heat flux information in this thesis will be the following literature review.

Ahrnsbrak (1968) assumed that the storage of heat in the soil above the permafrost would be negligible compared to the heat necessary to melt the permafrost. However, on the basis of laboratory determination of the water content of permafrost samples (30± 5 percent by volume) and the depth of permafrost below the ground (35-85 cm.), he decided that heat storage in the soil above the permafrost was not negligible. After some crude calculations Ahrnsbrak determined that for his wettest site the ground heat flux could be much larger than originally expected and account for almost 30 percent of the available net radiation. He was unable to take any samples of the permafrost at this site since any hole dug in an attempt to do so immediately filled up with water. It may be assumed therefore that the moisture content at this site was greater than the value obtained for the drier sites noted above. However since the wet site was composed of gravelly sand it is unlikely that Ahrnsbrak's wet site is comparable with the very high volumetric moisture contents associated with the sites investigated in the present study. The inference to be drawn from the study by Ahrnsbrak is that heat storage may account for a significant portion of the available net radiation, perhaps attaining 30 percent, though it is possible that the strong insulating properties of dried out moss may make this unrealistic during the late summer period.

Brown and Williams (1972) have presented some data on ground heat flow through *Sphagnum* covers at Thompson, Manitoba, which is situated similarly to Baker Creek basin in the middle of the discontinuous permafrost zone. The data were obtained using soil heat flux meters and although doubts are expressed as to their accuracy the results given appeared to be consistent with changes in the depth of the permafrost table during the period. At a site with a relatively thick *Sphagnum* and peat horizon, (35 cm.), and underlain by permafrost the mean summer ground heat flow averaged over two summers, (June-September inclusive; 1970,1971), was just over 2000 cals. cm.$^{-2}$. Over the same period at another site where the *Sphagnum*/peat layer was only 20 cm. deep the ground heat flow averaged just over 1600 cals. cm.$^{-2}$. Unfortunately net radiation measurements are not available thus precluding an assessment of the importance of these fluxes in terms of the total energy balance. It is interesting however, in anticipating the interpretation of the present study active layer development results, to note the detailed explanation given for the difference in flux between the two sites which had similar exposure and soil conditions. It was assumed that evaporation was almost similar at both sites since although one site is saturated the *Sphagnum* at the unsaturated site "acts as a wick providing ample water to the surface" (p19). Brown and Williams suggested that ponding of water at the large ground heat flow site affected the surface heat balance sufficiently to cause significant changes in total heat gained at the site.

The microclimate studies conducted at Point Barrow (Mather and Thornthwaite, 1958) resulted in some data on the relative importance of ground heat flux in an arctic energy balance. According to Nakano and Brown (1972) the thickness of the organic layer of the tundra soils at Barrow varies in the range of about 2-10 cm. before mineral soil is encountered. Monthly estimates of average daily heat balance data during the 1958 summer season when data were considered to be more accurate indicated that the proportion of net radiation used by ground heat flux was 15 percent, 13 percent, and 20 percent for June, July and August respectively (p 84).

Other data reported by Mather and Thornthwaite (1958, p 77) refer to the actual individual days of heat balance monitoring. Gill (1971, p 507) has calculated from these that 24 percent of the net radiation is consumed by ground heat flux on average during the main summer period. It may also be noted that in Gill's own study in the Mackenzie Delta he found that soil heat flux in the lower level of spruce stands during midsummer was about 27 percent of net radiation.

Addison (1972, 1973) has reported that ground heat flux through the sedge meadow site at Devon Island ranged from about 8-12 percent at midsummer for saturated conditions. Rouse (1973) has reported results for the Hudson Bay Lowlands at a lowland swampsite. This single day study indicated ground heat flux to be about 8 percent of net radiation. However Rouse also found that between mid-June and mid-August, 1972, the active layer beneath this site melted to an average depth of 35 cm. He noted that this process utilised in excess of 1500 cals. cm.$^{-2}$ of energy and was thus, "an important element in the heat storage of the subsurface layer" (Rouse 1973).

From this short review it would appear that ground heat flux constitutes a very variable percentage of net radiation. Some possible patterns may be identified however. It is likely to constitute a small proportion of the net radiation when sites are saturated to the surface but where the organic layer is relatively thin and underlain by a mineral soil where volume moisture contents are low. Where the organic layer is thick, moisture contents will be high especially if permafrost impedes drainage. The large amount of ice in the moss will require more energy for its melting. The water will also act as a reservoir of heat in the manner of a lake. If the surface is well insulated by a dried out moss cover for much of the summer the heat penetrating the cover may be low but the moss will also act to encourage storage of this heat in the manner of the greenhouse effect. If only part of the summeris considered which happens to be a period dominated by melting and warming in the ground it may be appreciated that ground heat flux could attain a significant proportion of the net radiation for this particular period.

LAKE EVAPORATION

Evaporation from lakes in Canada's North has received very little research attention and literature on the topic is scanty. This is regrettable when it is realised that lakes comprise a significant proportion of the total surface area in many parts of the North and hence constitute an important element in the water balance.

Semi-empirical estimates of free water evaporation in the study area may be extrapolated from nation-wide studies by Ferguson, O'Neill and Cork (1970) and Bruce and Weisman (1967). The monthly and total values for the study area as extrapolated from these studies is presented in Table 3-7. The estimates apply to free water surfaces with no heat storage effects and no advected energy influences. Because of the increased heat storage effect larger lakes can be expected to have somewhat less evaporation losses. The estimates were derived primarily from Class A evaporation Pan data. The concept at the root of the use of such pans for estimating natural evaporation has been criticised for being unsound physically (Thornthwaite and Hare, 1965). Allowance has to be gauged for additional heat for evaporation being supplied through the bottom and sides of the pan. Also the pan's rim creates unnatural turbulence effect. The most serious objection to the instrument however seems to be its approximation of 'mid-desert' conditions rather than the more desirable 'mid-ocean' state which would be partly met by locating the pan on a natural lake surface.

TABLE 3-7

ESTIMATES OF FREE WATER EVAPORATION (INCHES)
FOR BAKER CREEK BASIN BASED ON YELLOWKNIFE AIRPORT
MONTHLY AND ANNUAL TOTALS DATA

	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	TOTAL
Ferguson, O'Neill & Cork (1970, p 1627)	0.4	3.4	5.0	4.0	1.4	0.4	14.4
Bruce and Weisman (1967)	2.8	3.5	3.4	2.3	1.4	0.9	14.3

Present site controls on the pan network operation probably favour advective energy increasing evaporation to levels above that occurring naturally.

Ferguson, O'Neill and Cork (1970) note that in their 'northern' zone, which includes the present study area, "lake evaporation estimates must be viewed with some skepticism owing to the extreme paucity of evaporation measurements" (p 1629). The estimates are however the best available, incorporating several improvements over the earlier provisional maps produced by Bruce and Weisman (1967). The most notable improvement is the allowance for break-up and freeze-up influences. This improvement, in the study area at least, has tended to cause changes in the monthly evaporation regime rather than the annual total.

The more recent study involved the analysis of twice as many Class A pan data as the former study and additional climatic, (mainly sunshine), data. Four years of Class A pan data from Yellowknife Airport were used in the study; however on comparing the available pan data and the estimated values it appears that little attention was given to the existing records. The total values estimated tend to be well below those which may be taken off the available records (Table 5-5). Some indication of the lake evaporation patterns in the present study area may be gleaned from a recent study of the surface water temperature and evaporation regimes of selected lakes in Northern

Ontario (Richards and Webb, 1971). This study involved sensing of lake surface temperatures by an airborne infra-red-thermometer. Twenty-three surveys were completed in four years between break-up and freeze-up periods. Although a total of 35 lakes were monitored, only seven occur in the northern portion of the study area which is considered to be most similar to the present study area in terms of general climate and physiography. The general location of the seven northern lakes is between 52-54°N, 85-90°W (Fig. 3-1). The mean surface water temperatures by month and estimated mean monthly evaporation, including estimated annual totals, are presented in Table 3- 3.

The monthly evaporation was estimated from a mass transfer equation developed in the Lake Hefner and Lake Meade Studies:

$E = 0.0024\ (e_s - e_{ae})\ V$ where E is evaporation in inches per day; e_s is saturation pressure in inches of mercury at the surface water temperature; e_{ae} is the vapour pressure in inches of mercury, (obtained from climatotogical data); V is wind speed in miles per day, (obtained from climatological data. The annual evaporation totals incorporate a subjectively estimated amount, made by this researcher, for early or late season evaporation.

It may be observed that the larger lakes (Big Trout and Mississa) average about 10 inches annual evaporation while the smaller lakes average about 13 to 14 inches. The latter range corresponds very well with the estimate

TABLE 3-8. MEAN MONTHLY LAKE SURFACE TEMPERATURE AND EVAPORATION, NORTHERN ONTARIO LAKES.

N.B. Values in brackets are personal estimates,

Source: Richards and Webb, 1971

T Mean Lake Surface Temperature (° C) E Mean Lake Evaporation (inches)

Lake	Area (sq.mi.)	Latitude		May (29th.)	June	July	Aug	Sept	Oct	Nov (5th.)	TOTAL
Bug L.(west arm)	1.6	53° 37′	T.	6.3	11.7	19.0	14.6	11.5	6.4	0.7	
			E.	(1.11)	2.73	3.98	2.22	1.87	2.09		14.00
Missamikwash L.	3.2	52° 59′	T.	7.3	11.3	17.7	14.8	12.3	7.1	0.7	
			E.	(1.45)	1.66	2.94	1.88	1.84	2.03		11.80
Assinne L.	3.3	52° 50′	T.	8.4	13.0	18.3	15.9	13.2	6.8	0.9	
			E.	(0.84)	2.45	3.13	2.37	2.27	1.96		13.00
Big Trout L.	250.0	53° 45′	T.	1.5	5.5	14.4	14.9	12.4	6.3	2.8	
mean depth	52 feet		E.	0.18	1.29	2.18	2.25	2.12	(1.08)		9.00
Mississa L.	73.0	52° 19′	T.	3.4	9.8	18.0	16.3	11.9	5.7	0.4	
mean depth	5.7 feet		E.	(1.62)	1.87	2.34	1.93	1.53	1.71		11.00
Neawagank L.	4.9	52° 27′	T.	9.0	12.6	18.0	15.6	12.6	6.8		
			E.	(1.45)	2.20	3.09	2.30	1.95	2.01		13.00
Sandy bank L.	7.5	53° 50′	T.	5.2	11.4	18.0	15.0	12.3	6.9	0.5	
mean depth	3.6 feet		E.	(1.55)	2.56	3.20	2.13	2.22	2.34		14.00

of annual free water evaporation published for the area by Ferguson et al. (1970), which in turn is very similar to that mapped for the present study area (Ferguson et al, 1970, p 1626) (See Figure 3-1 for positioning of mean annual evaporation isolines).

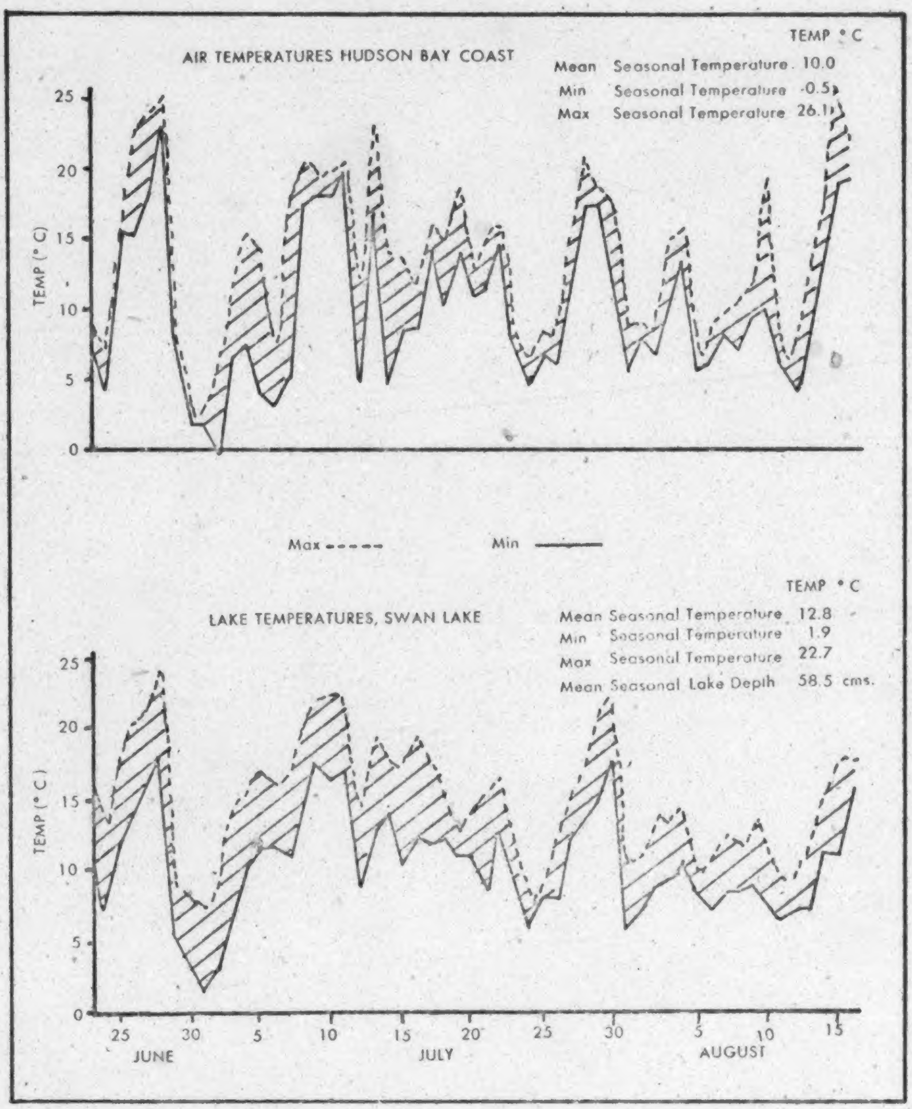
Despite these correspondences the limitations of the Northern Ontario study should be noted. The monthly means were calculated on the basis of a very small number of surveys--a total of 23 in four years. Other research in the same region, reported later in this section, indicates large fluctuations in lake temperatures from day to day and this characteristic would tend to subtract further from the accuracy of the results. The reliability of the equation used for calculating evaporation has not been proven experimentally for northern regions, and in addition it involves use of climatological data from stations which may be up to 100 miles distant from the lakes monitored.

Specific influences on lake surface temperatures across Northern Ontario, Eastern Manitoba and the Keewatin have been documented in detail by Peterson (1965), McFadden (1965), Webb (1973) and Rouse (1973).

Rouse (1973) includes some data on diurnal and daily temperature changes in a lake in the Hudson Bay Lowlands (Figure 3-3). It is evident that the lake undergoes a greater diurnal temperature change than the air above due to the strong absorption of solar radiation. At the same

FIGURE 3-3 COMPARISON OF AIR TEMPERATURES, HUDSON BAY COAST (LAT. 56° N) AND LAKE TEMPERATURES, SWAN LAKE. SUMMER 1972.

Source: Rouse (1973)



time the day to day temperature changes in the lake are almost as great as for the atmosphere. This lake is the one referred to in Table 3-5. In terms of the radiation balance the lower albedo of the lake is re-inforced by a lesser long-wave radiation loss so that its net radiation is substantially greater than over either the swamp or the ridge. Rouse indicated that the evaporation from the lake could not be compared directly with that from the other two sites because so much heat went into storage during the daylight hours. The lake would continue to give up heat during the night-time hours, much of it going into evaporation. Rouse found the energy budget of the lake interesting because the storage term was so large and responded to the advectional effects of warm and cold air masses as much as it did to radiational inputs. He noted the implication from Figure 3-3 is that there would have been a number of periods when the lake was cooler than the air above, leading to a downward directed sensible heat flux. Such atmospheric advection appeared to be very important in contributing to the highly variable day to day behaviour of the lake.

The freezing and thawing processes of lakes have been documented by Scott (1964) who studied the winter heat balance of many Wisconsin lakes. His conclusions are probably applicable to the present study area if accentuated by an order of magnitude. In late autumn, Scott noted that the lakes cooled rapidly by a transfer of sensible and latent heat to the atmosphere with the deep lakes losing heat more rapidly than the shallow lakes, and the temperature of the shallow lakes decreasing most rapidly. During this period Scott calculated an evaporative heat loss for large lakes of approximately 0.5 cms./day. This value is probably too high to be applicable in the present study area.

The wastage or ice melt period is characterised by both radiation gain and transfer of sensible and latent heat from the air to the ice. Once snow cover has disappeared from the ice surface an obvious greenhouse effect occurs. Roughly 60-75 percent of the effective solar radiation at ice surface is transmitted through the ice, is absorbed by, and heats the water. Conduction of heat from the water to the ice causes melting at the bottom of the ice and transfer of sensible and latent heat from the warmer air above results in wastage at the upper surface of the ice. The amount of heat available for melting the ice from below is a function of the size and depth of the lake. Radiation may account for approximately 25 percent of ice wastage from the lower surface in deep lakes and up to 50 percent of wastage in shallow lakes (Scott, 1964).

The apparent importance of solar radiation input in contributing to rapid warming of the shallow lakes in particular appears to be dependent on the nature of the ice cover (Billello, 1968; Williams, 1969; Jones 1970).

Billello (1968) has noted that solar radiation can contribute a substantial amount of heat to the water beneath an ice cover when the ice is clear. If this is not the case more than one half of the total incoming solar radiation is reflected from the snow and ice surface. Billello affirms that: "in any case most of the energy in the infra-red wavelength is absorbed in the surface layer. Most if not all of the energy absorbed is due to light scattering in snow-bubbly ice layers (Billello, 1968, p 754).

In his study at Seneca Lake, northern Michigan, Billello observed that water temperatures underneath the ice cover in the shallow lake decreased substantially during the ice thaw period. Solar radiation appears to have had little heating effect by comparison with the cooling influence of snowmelt runoff penetrating below the lake ice cover. The decreased influence of solar radiation in Billello's study is probably related to the fact that two-thirds of the lake ice cover was found to be bubbly.

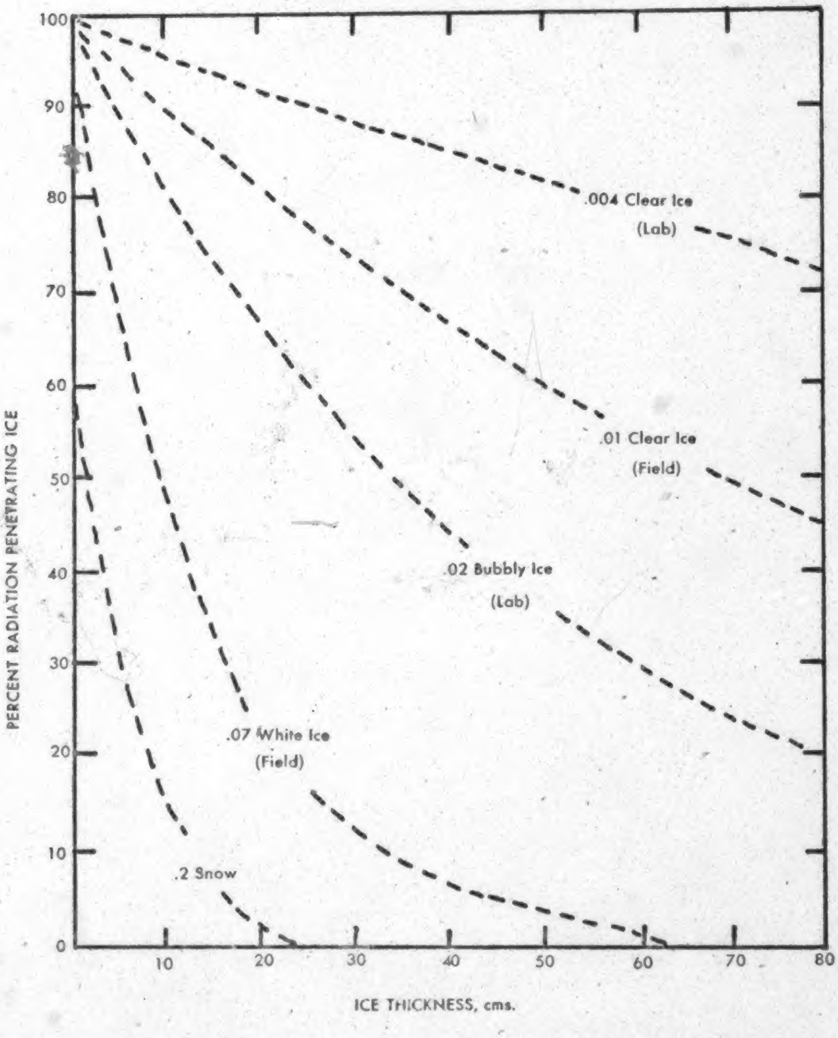
Williams (1969) found contrasting results for a similar study at White Lake near Ottawa. He observed that water temperatures under the ice cover increased substantially during the thaw period with temperatures as high as 7.5°C being recorded. Calculations indicated that solar radiation penetrating the ice caused the relatively high observed temperatures. The ice on White Lake was essentially clear but had some air bubbles entrapped in the upper layers. Williams estimated that the extinction coefficient was in the range 0.01-0.02. He also constructed a diagram, from reports in the literature, relating ice thickness, percentage radiation penetrating the ice and extinction coefficients

(Figure 3-4). This shows that a large proportion of incoming solar radiation will penetrate clear ice of considerable thickness, whereas the amount penetrating bubbly ice or snow-covered ice may be almost negligible.

White ice was found to be an important component of the ice sheet in the Knob Lake area of Central Labrador-Ungava (Jones, 1970, p 119). Jones felt that it was likely to be a common element of all lake ice covers in subarctic regions generally. He noted that it directly encourages ice growth. The main controls of white ice growth appeared to be snowfall and high diurnal temperature ranges. In Canada such conditions are best fulfilled within the taiga section of the Subarctic, where precipitation is well-distributed and several months with below freezing temperatures exist. It is possible that in the present study area the much lower precipitation, in addition to local drifting patterns, will result in a less intense white ice development. White ice certainly appears to be a significant element in the Knob Lake area. Hare and Hay (1971, p 88-9) have noted in this region that lake evaporation falls well short of that possible according to the net radiation input. This is explained by the influence of the snow covered (white ice protecting the lakes most effectively from intense season solar radiation. Penn (1970) has provided further evidence of this effect in the same region from lake temperature profiles. The ratio of the energy used in small-lake evaporation to the net radiant income at Knob Lake is 0.60 whereas for the present study area it is 0.80 (Hare and Hay,1971,p89).

FIGURE 3-4 PERCENT RADIATION PENETRATING ICE COVERS FOR DIFFERENT EXTINCTION COEFFICIENTS

Source: Williams, (1969,p.1137)



Where early season lake surface temperatures may lag significantly behind air temperatures the possibility of condensation onto lakes must be considered. The largest body of literature on this aspect in Canada relates to the Great Lakes (Richards and Irbe, 1969). Condensation gains are significant during spring and early summer for the Great Lakes. The slow warming of the large water bodies delays evaporation until late summer and early winter. Lake Superior, the northernmost lake, has later condensation and evaporation seasons due to its particularly slow warming pattern.

Laycock suspects that condensation may be significant for the lakes in the present study area, particularly for the deeper lakes (Personal Communication, December 1971). Webb has commented on this possibility in the light of his northern Ontario study (Webb, Personal Communication, May 3rd, 1973). He points out that for a lake to gain moisture through condensation the advected air must have a dew point temperature which is warmer than that of the ice surface. For small, shallow lakes in Northern Ontario mean monthly surface temperatures were found to be warmer than the corresponding air temperatures. Since dew point temperatures are several degrees cooler than air temperatures, the occurrence of condensation would be infrequent. Webb notes however that for deeper lakes the frequency would be greater, particularly during the spring warming when these lakes would be cooler than nearby shallow ones. He has compared mean water surface temperatures for Big Trout Lake and mean dew point temperatures from the Trout Lake weather station. (Table 3-9).

TABLE 3-9

COMPARISON OF MEAN WATER SURFACE TEMPERATURES
AND MEAN DEW POINT TEMPERATURES,
BIG TROUT LAKE, NORTHERN ONTARIO

	Jun	July	Aug.	Sept.	Oct.
Dew Point Temperature °C	6.1	11.1	9.4	5.0	-0.6
Surface water temperature °C	7.2	14.7	15.7	12.1	6.5

Source: Webb (Personal Communication, May 3rd, 1973).

Webb commented:-

> "These data indicate that condensation into this deep lake is most likely to occur in June. Even then the quantity of moisture which could be gained on these occasions would be limited by the temperature inversion which must exist over the lake. For these reasons it if felt that the potential moisture gain by this lake from condensation could be neglected in any water balance computations for this region. I feel the same conclusion would apply in the Yellowknife area".

It may be added that Big Trout Lake is very much larger and probably has a greater mean depth than the largest lake in the present study region, excepting of course Great Slave Lake.

An estimate of lake evaporation patterns in the present study has been provided by Kakela (1969, p 235-6), for Pocket Lake. His mean annual water balance for this site reduced to an input of 12.45 inches precipitation, 15.04 inches evaporation, and the lake level was maintained by 2.59 inches of riparian runoff into the lake from the adjacent land surfaces. A slight loss of snow from the lake surface by drifting was included in his winter site-seasonal water balance equation. Kakela believed the lake surface to be slow to heat in spring because of the initial need to melt the lake ice and the low elevation of the sun in spring causing a high proportion of the incoming solar radiation to be reflected from the water surface. This led Kakela to lower the spring season P.E. estimate for the lake by comparison with most of the land surface sites. Kakela also felt that the lower spring rates of P.E. were not compensated for during the summer months but that the air layer over deep lakes, (eg Pocket Lake), tended to have a slightly cooler thermal regime than adjacent land surfaces. This he felt would be mainly caused by the 'marine' effect of the deep water bodies and the short summer heating period in the Yellowknife area. As a result he reduced the mean annual potential evaporation by 1.5 inches by comparison with the P.E. used for land surface equations (Kakela 1969, p 236). Kakela makes no mention of the thickness and nature of the ice cover on the lake. This would seem to be a useful area for future study in view of the previous comments on the interrelationships between ice cover characteristics, solar radiation receipt, and early season warming patterns.

BASIN-SCALE EVAPOTRANSPIRATION

In the absence of measured values, working estimates of evapotranspiration may be derived from empirical models, net radiation or free water evaporation considerations, or from climatonomic analyses.

The most widely applied empirical water balance model is that developed by Thornthwaite (Thornthwaite 1948; Thornthwaite and Mather 1955; 1957). Numerous successful applications of Thornthwaite's model have been recorded for a variety of mid-latitude locations, mainly in North America (Mather 1954). The unqualified applicability of the model in climatic environments outside the mid-latitudes was questioned from the outset (Thornthwaite, 1948, p88). These doubts seemed to have been removed for Northern Canada locations as a result of studies by Sanderson (1948a, 1948b, 1950). However these indications of early support for the unadjusted Thornthwaite model have been undermined since as a result of intensive experimental investigations considering indigenous vegetation, and also through comparative studies of model-computed moisture surplus and the gauged runoff values that have become available for many northern river basins in the last few decades (Mather and Thornthwaite 1956; 1958; Brown, 1961; 1965; Kakela 1969; 1973; Cavadias, 1965; Hare 1955; Hare and Ritchie, 1972, p 354).

Many of the criticisms of the model for northern areas become less serious when only the mean annual patterns are considered rather than monthly or daily patterns. Kakela

(1973) has shown that the major deficiencies of the model can be related to the questionable bookkeeping assumptions made in the procedures associated with the model. On an annual basis many of these deficiencies are removed. In particular Kakela indicated that varying P.E. rates in an attempt to make them more representative of the latitude under consideration had virtually no effect on residual actual evapotranspiration and runoff values. There is therefore some justification for using mean annual values derived from this model as working estimates in the present study.

Potential Evapotranspiration (P.E.) is the key concept in the Thornthwaite water balance model and has been defined as the water loss from a moist soil tract completely covered by vegetation (without specifying the type), and large enough for oasis effects to be negligible (Thornthwaite and Hare 1965,p 168). Since moisture is not restricted, P.E. is limited solely by available energy. As long as moisture is abundant P.E. is taken to depend primarily on meteorological conditions with variations in natural vegetation and soil type having little influence on water losses. Computation of P.E. may be accomplished by knowing only the latitude and mean temperature of the location for which an estimate of the water balance is desired. The expression relating these factors is in effect a regression of evaporation from certain well-watered U.S. drainage basins and lysimeters upon monthly temperature. As a result of the Point Barrow, Alaska, studies the validity of this regression for calculating P.E. for high latitude surfaces was questioned. Subsequent heat balance studies in northern areas have also indicated that the compartmentalization of the energy budget there differs from that assumed in the model. It is therefore in order to view the following P.E. estimates as probably requiring some downward adjustment.

A critical factor in the Thornthwaite model affecting actual evapotranspiration and runoff values is the assumption related to soil moisture storage capacity. It is very difficult to estimate this value for northern areas. Kakela (1973) suggested that if measured discharge values are compared with calculated runoff values for a variety of different storage capacities it may be possible to obtain an estimate of mean storage capacity for the general area drained by the river for which measured runoff is known. He has provided a graph of such relationships for rivers in the present study region (Kakela, 1973, p174). The reliability of this approach for estimating storage capacities depends to some extent upon the accuracy with which the Thornthwaite climatic water balance describes the actual hydrologic situation in the area. The relationship should therefore be used with caution for estimating representative average storage capacity values.

The water balance predicted by the model assumes that the initial precipitation input is accurate. Many studies exist indicating the unlikelihood that this assumption is valid for northern areas which receive large amounts of snow precipitation which is usually undermeasured (Wilson, 1954; Hare and Hay, 1971; Findlay 1969). However this deficiency has to be faced in any attempt to define northern water balance patterns.

Brandon (1965, p 7) computed monthly P.E. using the Thornthwaite method for Yellowknife. The results are presented in Table 3-10.

TABLE 3-10

AVERAGE MONTHLY P.E.,

YELLOWKNIFE AIRPORT

	May	Jun	Jul.	Aug.	Sept.	Year
P.E. (inches)	1.6	3.7	4.8	3.9	2.2	16.2

Source: Brandon, 1965, p 7.

M.E. Sanderson has contributed much of the effort in applying Thornthwaite's model to Canada. Recently she has summarised this work, (Sanderson, with Phillips, 1967; Sanderson 1969). She indicates that the water balance patterns mapped are based primarily on the water surplus data published in a C.W. Thornthwaite Associates manual, (Mather, 1964). In addition however she used 1931-1960 climatic normals for 150 stations. Interestingly in the case of Yellowknife use of the latter data resulted in an 18 year mean annual precipitation of 10 inches compared with the Thornthwaite namual normal of 8.4 inches (10 year mean). Gauged runoff data for the Yellowknife river was also utilised in constructing the maps. A 21 year mean runoff of 2.8 inches was noted.

Some tentative estimates of the water balance parameters for the Baker Creek basin may be extrapolated from Sanderson's maps. Average annual P.E. for the basin is indicated as being almost centrally located within the 11.22 to 16.83 inches class. Though the poor quality of the maps makes it difficult to definitely pinpoint surplus and deficit values for the area there are indications of a very limited surplus in the range of zero to two inches and a deficit in the range of four to six inches. It ~~may~~ be of interest to note that for the headwater areas of the Yellowknife River a surplus of about four to eight inches is indicated. All these values relate to a soil moisture storage assumption of 12 inches.

On the basis of Sanderson's (1969) estimates an approximate water balance equation for the Baker Creek area may be constructed as follows.

* $P = (PE - D) + S \pm S.C.$

(inches) $10 = (14 - 5) + 1 \pm 0$

It must be stressed that this equation is based on a strict interpretation of values directly extrapolated from relatively poor quality maps. This explains the discrepancy between the P.E. value selected here, (14 inches), and the value computed from Yellowknife Airport data, (16 inches), used in the studies mentioned above. In spite of this discrepancy it was possible to obtain a water balance quite easily. The interpretation is presented here primarily because it is the closest P.E. - based estimate to the actual water balance that may be deduced from the present research for the vicinity of Yellowknife Airport.

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* P = Precipitation P.E. = Potential Evapotranspiration.
D = Deficit S = Surplus S.C. = Storage Change.

Laycock (1968) commented on the Sanderson and Phillips study (1967), with particular reference to Yellowknife. From the study he interpreted the approximate water balance equation for Yellowknife to be:

$$P = (P.E. - D) + S + S.C.$$

$$\text{(inches) } 8.5 = (16 - 7.5) + 0 \pm 0$$

(Laycock, 1968, p 7).

Laycock proceeded to outline a number of reasons why streamflow should be greater than zero:

> "Much of the surplus occurs in the wetter than average months of some years and this would not be shown if average monthly data only are used. Precipitation measurement, particularly of snowfall, is not as accurate as we might wish--the efficiency of catch may be 10 to 40 percent below actual receipt, and this appears to more than compensate for errors in conversion of snow measurement to water equivalent relating to the low density patterns that are more widely recognized. We need precipitation measurements for some upland areas, but snow course and other measurements near existing stations could help us to correct records we now obtain. The potential evapotranspiration estimates appear to be too high for most areas. The high albedo of snow and ice, present until early June on the large lakes, and of some light-toned rocks and sedge meadow areas and the cool surfaces of the larger lakes, and of soils over permafrost probably reduce potential evapotranspiration from the levels indicated. Actual evapotranspiration is lower than that indicated for large areas--e.g. bare rock areas, which are widespread in the North, have rapid runoff and only a few inches of the precipitation may be evaporated or transpired.

> Snow drift areas may have significant local surpluses whatever the soil and cover patterns present. For these and other reasons some of the small basins near Yellowknife apparently have average yields of several inches" (Laycock, 1968, p 7).

The most detailed application of the Thornthwaite water balance procedure in the Baker Creek basin area may be attributed to Kakela (1969, 1973). On the basis of fieldwork in Pocket Lake sub-basin Kakela selected a one inch soil moisture storage assumption for his calculations and he also assumed that 50 percent of the available surplus water per month was manifested as streamflow.*

The 25 year mean annual water balance calculated by Kakela for Yellowknife Airport data is presented in Table 3-11. P.E. was calculated on the basis of the 50°N latitude correction factor provided in the Thornthwaite computation tables. Kakela developed a correction factor for 62°N, Yellowknife's latitude, and the water balance using this method is also presented in the table.

TABLE 3-11

CALCULATED 25 YEAR MEAN ANNUAL WATER BALANCE FOR YELLOWKNIFE AIRPORT FOR TWO LATITUDINAL CORRECTION FACTORS.

Ppct = (P.E. - D) + S ± S.C.

62°N correction 9.7 = (18.5 - 12.1) + 3.7 ± 0

50°N correction 9.7 = (17.3 - 11.4) + 3.8 ± 0

Soil moisture storage capacities = one inch

Source: Kakela, (1973 p 171, 175)

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* Sprinkler experiments over surfaces in the basin have indicated that the 50 percent value may be too low (Laycock, Personal Communication, August 1973).

The calculated runoff, which remained essentially the same whatever the correction factor used, was almost similar to the measured 18 year mean discharge for the Snare River (3.9 inches) but greater than 26 year mean discharge for the Yellowknife River (3.0 inches).

Numerous attempted correlations between calculated runoff and measured runoff for individual years failed to produce any significant results. Some success was obtained however from attempts to correlate calculated runoff and certain precipitation variables. Specifically, there was evidence that surplus precipitation is delayed (detained) for an unusually long time period before it materialises as stream discharge from the larger drainage basins of the Yellowknife area. There was also an indication that summer precipitation (rain) received during the 12 months preceding discharge could be contributing a significant portion of the surplus generated in the area (Kakela, 1969, p 221). The latter correlation was the highest obtained but it could not account for more than 35 percent of the variation of stream discharge from year to year (Kakela, 1973, p 175).

Kakela (1969) developed numerous site - seasonal water balances for Pocket Lake basin (p 227-237), focussing on the snow aspects of the water balance and the winter situation for nine different site types. (These were refined as a result of the present research and are presented in Chapter 6). On the basis of these, Kakela calculated the mean water balance for the entire sub-basin by an area weighting procedure. The resulting water balance equation was as follows:

$$P \quad = \quad (P.E. \; - \; D) \; + \; S \pm 0$$

$$(\text{inches})\,12.76^{*} = (16.00 - 6.28) + 3.04 \pm 0$$

(Kakela, 1969, p 237)

The area weighted mean precipitation and surplus values are virtually identical to the adjusted mean precipitation and mean measured discharge values, respectively, that were recorded in the Yellowknife area. Kakela admits that the estimated water balances for the specific sites were manipulated so that these values would result. However, Kakela (1969) believed that more accurate estimates of mean P.E. and deficit, (and hence actual evapotranspiration), were determined when compared to the values calculated for the Yellowknife area using the selected one inch soil moisture storage capacity.

Net radiation values are an another potential source of working estimates of basin-scale evapotranspiration. According to Sanderson (1969, p45): "It is generally accepted that net radiation is the parameter which could most accurately be used to estimate P.E.".

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* Kakela adjusted the precipitation values on the basis of correction factors derived from a Soviet review article. U.S.S.R. Interdepartmental Committee for the I.H.D., "Summary of Methods of Computation of the Water Balance", February 1967, 19 pp (Mimeographed). As a result of the review conclusions, and his knowledge of the specific Yellowknife area characteristics, Kakela suggested increasing the 25 year mean snow precipitation as gauged at Yellowknife Airport met. station from 9.74 to 12.75 inches. This was based on an upward adjustment for snow precipitation of 50 percent and rain precipitation was increased by 15 percent (Kakela, 1969, p 226).

Net radiation has been measured at Yellowknife Airport for a period of three and one-half years between August 1960 and November 1963 (Wendland, 1967). The instrument used was a Suomi-Kuhn net radiometer, the accuracy of which is only within about ± 10-15 percent. Further error is introduced into the measurements by beam radiation reflection off the polyethylene surfaces of the instrument. The range of error for this phenomenon varies up to 15 percent of measured values depending on cloud cover conditions. Wendland has indicated elsewhere that for regional representativeness his measured results should be increased by 35 cal. cm.$^{-2}$day^{-1} * (Wendland, 1965). This suggestion is followed in the presentation of his results in Table 3-12.

Other estimates of net radiation are available for Yellowknife, but they are empirical in nature (Hay, 1970; Budyko, 1963). Both these workers have employed computational models, using standard climatological data to compute the radiative fluxes for a particular location. Their results are also presented in Table 3-12. The table includes an estimate of the annual potential evaporating power of the net radiation totals from each source.

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* This adjustment factor was intended to account for differences between the net radiation measurement site at the Airport, (sandy, gravelly beach ridge with a few grass tussocks; relatively high albedo), and the surrounding Boreal Forest, (relatively low albedo).

TABLE 3-12

NET RADIATION ESTIMATES FOR YELLOWKNIFE, N.W.T.

(Daily Mean sums of net radiation per month)

(calories per square centimeter)

	J	F	M	A	M	J	J	A	S	O	N	D	YEAR
Wendland (1965)	10	5	-5	105	300	335	295	215	115	45	20	25	123
Budyko (1963)	-33	-32	-15	33	139	277	226	161	83	10	-33	-33	66
Hay (1970)	-51	-45	-21	68	229	296	249	153	45	-16	-41	-50	69

Source: Hay, 1970, Table 6.31, p 160.

ANNUAL POTENTIAL EVAPORATING POWER OF NET RADIATION

	Annual Net Radiation (Kly)	cms.	ins.
Wendland	44.9	76.1	30.0
Budyko	24.1	40.8	16.1
Hay	25.2	42.7	16.8

Sources: Referenced literature and calculations by the author, where latent heat of vaporization was assumed to be 590 calories per gram

It may be noted that the empirical estimates of Budyko and Hay are compatible with one another but both are significantly lower than Wendland's measured estimates. Hay (1970) has attempted to explain the discrepancy by reference to Wendland's statement that the record of the economic radiometer was not continuous since it had to be covered during periods of precipitation and strong wind. This fact led Wendland to omit "many five to six hour segments---due to weather conditions " (Wendland, 1965, p6, quoted in Hay 1970, p 162). Hay felt that it was quite possible that during such periods of inclement weather net radiation values were lower than normal and hence the omission of such periods would tend to bias the measured monthly mean values to produce overestimates. He quoted evidence from Robinson (1964) to suggest that this would indeed be the case (Hay, 1970, p 162).

According to Hare and Hay, (1970, p 83) the assertion that such radiation estimates are the best available estimates of P.E. is controversial. They have noted that Budyko asserts that this assumption is valid only if the net radiation is derived for a fully moistened surface, (Budyko, 1968). Hay's estimates refer to actual conditions of moistening. They also feel it is probable that true equilibrium evaporation (after an infinite fetch across a moistened surface) is substantially less than the values computed by Hay. These comments in effect restrict the usefulness of the net radiation values as an estimate of

P.E. This usefulness is further reduced for seasonal and monthly estimates by having to allow for energy required for snowmelting, or warming the soil and lake waters.

Additional complications in northern locations have been documented by Orvig (1961). He attempted to correlate net radiation with measured evapotranspiration from a lichen covered evapotranspirometer at Knob Lake. No significant correlation could be obtained and Orvig attributed this fact to the peculiar characteristics of the lichen cover. Due to these characteristics he concluded that net radiation flux measurements "are of little or no help in evaluating evapotranspiration " (Orvig, 1961, p 203).

It may be concluded that the best use of the net radiation estimates may be in the derivation of actual, rather than potential, evapotranspiration estimates by means of Bowen ratios which reflect the peculiar characteristics of the major northern vegetation and surface types. This aspect is discussed more fully in chapter five.

According to Ferguson, O'Neill and Cork (1970, p 1631) there is a consensus that free water evaporation estimates can provide a useful index of P.E. over most of Canada. They indicate that in relatively humid areas there is support for the thesis that lake evaporation is similar in magnitude to actual evapotranspiration (p 1632). In more arid areas actual evapotranspiration tends to be limited not by the energy available but rather by water supply.

It is difficult to classify Baker Creek basin in arid and humid terms. Under the Thornthwaite climatic classification the region may be classified as sub-humid. Although precipitation levels indicate a semi-arid environment water supply is relatively generous because of the widespread lake and bog surfaces. It is probable however that the even more widespread bedrock-dominated surfaces limit water supply for evaporation to levels approximating somewhat less than the mean summertime precipitation for these surfaces. It is unlikely therefore that the mean free water evaporation values for the present study area (Table 3-9) constitute very good estimates of basin actual or potential evapotranspiration. The values are probably closer estimates of integrated lake surface evaporation only, although interpretation of the results of the present research indicate the estimates are on the low side.

A recent study, adopting a climatonomic approach, has provided further working estimates of evapotranspiration for the study area. (Hare and Hay, 1971). The study was confined to the large-scale annual water balance of northern North America, and involved a comparison of the best available measurements of precipitation and runoff in the light of the new estimates by Hay (1970) of the net radiation income of the surface. In the study an attempt was made to balance the heat and water budgets by focussing on the evapotranspiration component common to both. Anomalies were discovered suggesting that the mean fields of annual precipitation and runoff were inconsistent, particularly in northern Canada.

Hare and Hay (1971) were able to represent this inconsistency quantitatively by means of climatonomic runoff and dryness ratios. Lettau (1969), drawing heavily on Budyko's work on the world's heat balance (Budyko, 1958) has established a semi-empirical relationship between the two ratios. Hare and Hay used this relationship as a form of datum for judging the adequacy of the regional water balance patterns derived from currently available measured data.

The runoff ratio C is equal to the quotient of runoff to precipitation. The dryness ratio D is equal to the quotient of net radiation to heat equivalent of the precipitation (Latent heat of vaporization multiplied by precipitation). Annual means only are considered and no long-term changes in storage are assumed.

Regarding the relationship between the two ratios Lettau (1969, p 694) has written: "Supported by the analysis of actual precipitation, runoff, and net radiation data from a considerable number of watersheds, several possible but similar interpolation formulas have been suggested. Mathematically simplest and reasonably representative appears to be the following, expressed in ratios defined by annual means.

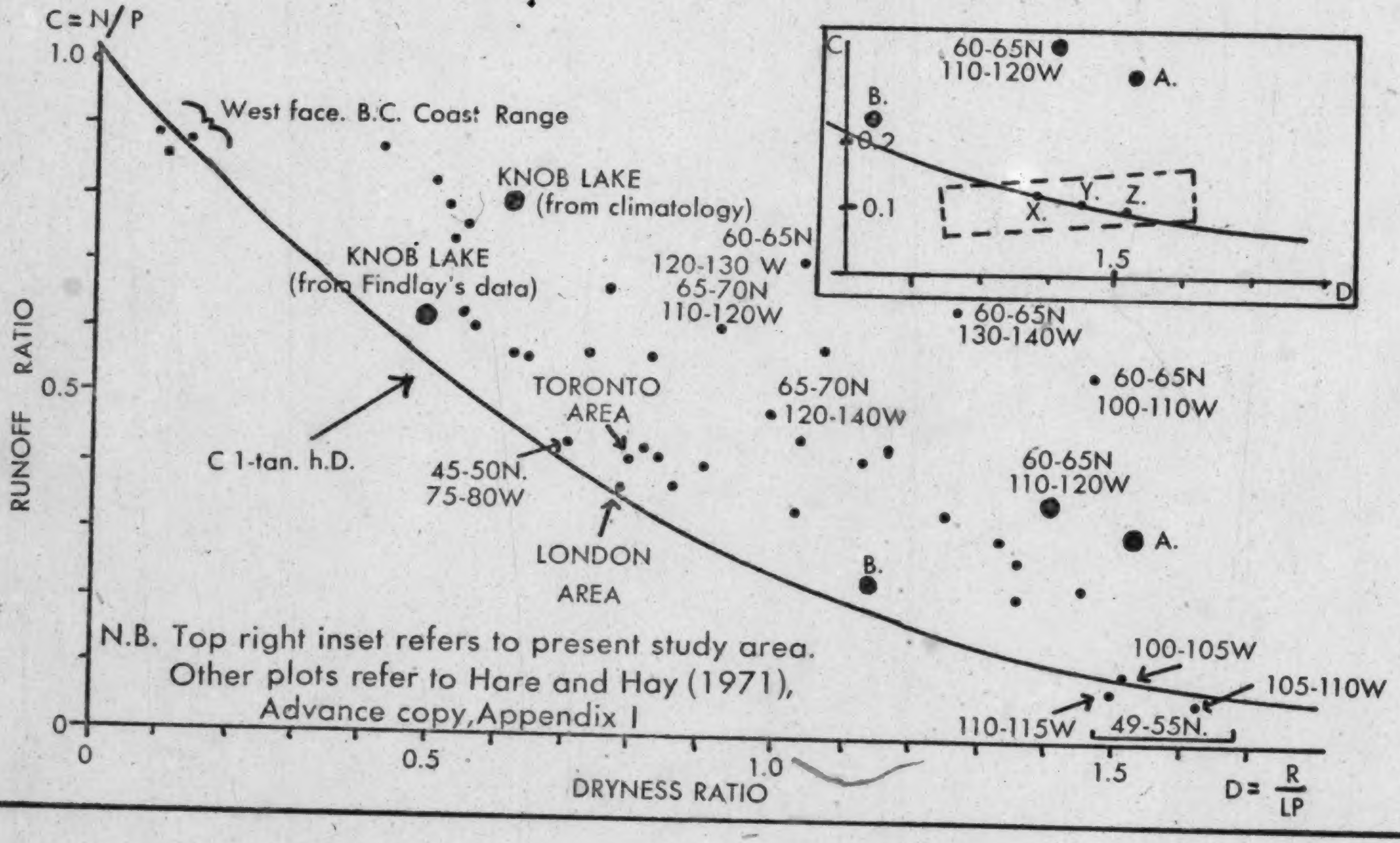
$$C = 1 - \tan h.\ D."$$

Hare and Hay (1971) generalized the various fields spatially and computed spatial mean values for various latitude-longitude squares, the basic unit being the five-degree square with a few modifications.

The plot of the runoff ratio against the dryness ratio for the squares used is shown in Figure 3-5. Also plotted is Lettau's function C = 1 - tan h.D. It is apparent that almost all the Canadian data lie above the curve of the function. Certain squares representing Northern Canada have been identified on Figure 3-5 and may be observed to have a particularly poor correspondence with the function. If Lettau's rule is accepted the misfit of the Canadian data can be explained by a systematic undermeasurement of precipitation, over-measurement of runoff or both. It is also possible that Lettau's rule fails in Canada since the curve is a semi-empirical function derived with little or no distinctly Canadian input. However, several observations point to its validity. Areas plotted very close to the curve are almost exclusively those possessing the highest quality of water balance data, and they also cover a range of climatic and physiographic regions eg. the more populated parts of Ontario, Quebec, the Prairies, and the West Face of the B.C. Coast range. In addition Findlay's subarctic study at Knob Lake (Findlay, 1966; 1969), provides strong evidence favouring validity of the function. His intensive measurements revealed large undermeasurement of precipitation, particularly of snow precipitation. As a result of his work, which happened to be conducted in an area where the previously available data caused a particularly poor fit with the curve, the fit was much improved. Since the function therefore appears to be valid the extreme distance

FIGURE 3-5: SPATIALLY SMOOTHED RUNOFF RATIO AS A FUNCTION OF DRYNESS RATIO.
Curve shows Lettau's empirically derived relationship between the ratios.

Sources: Hare and Hay (1971,p.90); Lettau (1969).



N.B. Top right inset refers to present study area.
Other plots refer to Hare and Hay (1971),
Advance copy, Appendix I

of the plotted northern areas from the curve can be taken as indicative of gross deficiencies in the data presently existing for northern water balance parameters. It is the purpose of the present study to attempt to remove these deficiencies for Baker Creek basin through a better understanding of the water balance patterns.

In their paper Hare and Hay (1971) stress the deficiencies in precipitation measurement, particularly snow precipitation, over much of northern Canada, as a major contributing factor to the anomalies between the heat and water balances. They quote Russian literature which has justified an upward adjustment of about 40-50 percent for the climatological normals of precipitation for northern parts of the U.S.S.R. (Bochov and Struzer, 1970). Hare and Hay dwell much less upon possible runoff measurement unrepresentativeness which was found to be significant in improving the water balance of the present study region.

The broad method of Hare and Hay (1971), and particularly the use of the Lettau function, can be employed in the present study to guide and test water balance manipulations aimed at deriving better evapotranspiration estimates. In Figure 3-5 the study area lies within the grid located between 60-65°N and 110-120°W. The plot for this grid area lies well off the curve of the function.

The plot is based on the spatially smoothed grid square values of net radiation, precipitation and runoff. These values differ slightly from the long-term means for the present study area, namely--Yellowknife Airport precipitation and Yellowknife River measured runoff. When these values are used the plot results in a slightly better fit but a large discrepancy with the curve remains. It is apparent that many plots could be generated by varying the precipitation and runoff variables in accordance with the results of more detailed research into local water balance patterns. The function may thus serve as a hypothesis-generating and hypothesis-testing device for working estimates of basin evapotranspiration.

CHAPTER IV

RESEARCH METHODS, STUDY SITES, AND LIMITATIONS OF THE STUDY.

INTRODUCTION: In this chapter the special characteristics of the area where research was concentrated are stressed, a description of the main study sites is given and the general limitations of the study are outlined. Methods of obtaining data for individual field projects are described in addition to some comments on their associated limitations.

GENERAL CONSIDERATIONS

Research in the present study was concentrated in the southern portion of Baker Creek basin, approximately four miles from Yellowknife and twelve miles from the head of the basin. It is in this region that the incidence of fires and chemical pollution has been greatest. It is not possible to evaluate the past and present implications of these factors for the water balance, but they should be borne in mind.

The major attraction of the basin from a research point of view is the relative wealth, by northern standards, of climatological data for the immediate area. Yellowknife Airport meteorological station may be taken as representative for many climatic parameters for the southern portion of the basin, though it would be most useful to have a detailed study on its exact degree of representativeness. Yellowknife Hydro climatological station appears to be well located to give a good representation of the climate and weather patterns for the headwater region of the basin. However this station is at an elevation (523'a.s.l.) approximately 200 feet below that for the upper basin and it is in fact lower than Yellowknife Airport (682'a.s.l.). In addition the quality of the snow precipitation records is suspect since the "ruler method" is employed. As will be shown later this method may well tend to result in an over-estimate snow precipitation for this station. In the present study therefore the record of Yellowknife Airport station has been used.

This study has benefitted greatly from the availability of runoff records for the much larger Yellowknife river were the only ones available for use in water balance definition. It had to be assumed that Yellowknife Airport records were representative of an area of over 4000 square miles - a most unlikely situation since bioclimatic gradients appear to change particularly rapidly in this region (Hare and Ritchie,1972). The lack until 1968 of locally representative runoff records has contributed greatly to a less refined interpretation of the actual water balance of the immediate area surrounding Yellowknife than is now possible (Kakela, 1969).

MAIN STUDY SITES

The major field research in the present study was associated with determining the evaporation or evapotranspiration from certain vegetation covers. Three main sited were studied and may be identified as YT2, YT4 and

4

YT6 in figure 1-3 (p4). The symbols refer to the ground temperature sites for the permafrost study conducted jointly by Dr. R.J.E. Brown, NRC, and Dr. A.H. Laycock, Department of Geography, University of Alberta. They may be elucidated as follows with approximate elevations in feet included.

YT2 - Muskeg (650)
YT4 - Sedge Meadow (660)
YT6 - Black spruce woodland bog (675)

These symbols are retained for the remainder of this thesis.

The muskeg site, YT2, is located between Pocket Lake and the stream course of Baker Creek. There is a slight, but perceptible slope in the direction of the Creek (Plate 4-1). The site has been well wooded at one time but suffered burning about ten years ago. As a result many upstanding barkless 'snags' or dead branches cover the site. This has contributed to higher albedo patterns than may normally be associated with muskegs since new growth of young trees and shrubs is presently not sufficient to shade the light-coloured deadfall.

The ground surface is made up of moss hummocks. (Plate 4-2). The dominant moss appears to be a Bryum (spp.), though some Polytrichum commune is present. The moss organic layer is relatively thin by comparison with the other sites being only about six inches in depth (Plate 4-3). Below this there is a sandy, silty mineral soil with some clay lenses suggesting a glacio-lacustrine origin. No permafrost is found at YT2, therefore when the seasonally frozen zone

Plate 4-1: View of YT2 (Muskeg). Pocket Lake off to right, Baker Creek to the left. Note the amount of deadfall littering the ground surface and the slight slope towards Baker Creek. Photograph taken facing in an easterly direction.

Plate 4-2: YT2 Evaporimeters battery. (10 cans).

Plate 4-3: YT2 Sod showing organic surface layer and the top of the mineral soil. See text for details of site characteristics.

Plate 4-4: View of YT4 (Sedge Meadow) David Lake lies about 100 yards off to the left of the photograph. View is taken approximately in a southeast direction.

Plate 4-5: YT4 Evaporimeters battery. (10 cans).
Note shrub area in background.
See text description.

Plate 4-6: YT4. Sod, showing sedge and moss understory.
See text description.

thaws out this site becomes relatively well-drained.

Apart from the evaporimeter and soil moisture measurements the net radiation results used in this study were obtained at YT2. As will be argued later there is a good case for expecting this site to produce regionally representative net radiation values.

The sedge meadow site, YT4, is located two miles north of YT2, about one quarter of a mile along the Muskeg Creek trail from the Ryan Lake road. The site occupies a drainage route between David Lake and a smaller unnamed lake.* (Plate 4-4). Apart from a few isolated trees and clumps of shrubs the dominant vegetation at the site is sedge grass (Plate 4-5). Less noticeable, but very significant in evapotranspiration terms is the blanket of moss, Bryum (spp.), at the base of the sedge. In effect the stems of the _Carex_ punch through this blanket, the moss comprising approximately 70-80 percent of the ground surface. The moss grades into a dense organic layer, (Plate 4-6) which continues to a depth of at least 50 feet. This is due to the site lying immediately over the West Bay Fault bedrock depression.

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* The most probable source of water draining through the sedge meadow is David Lake. It is unlikely that the amounts involved are very significant since there is an intermittently flowing channel draining David Lake to the northeast. There is no well-defined drainage channel through the sedge meadow but only a string of widely-spaced small pools.

Plate 4-7: View of YT6 (Black Spruce Woodland bog). Muskeg Creek sub-basin off to the left of the picture; small lake about 50 yards off to the right. Photograph taken facing in a northerly direction. See text for fuller site description.

Plate 4-8: YT6M. Evaporimeters battery. (10 cans). Note the linear sedge pool area in background and the disturbed nature of the site.

Plate 4-9: YT6M. Sod showing compacted, relatively thin moss cover. See text

Plate 4-10: YT6S. Evaporimeters battery. (8 cans). Note shaded nature of the site. See text for fuller description of site characteristics.

Permafrost is found at YT4, the permafrost table at the end of the summer being at approximately 2.5 feet on average. The range in depth of the active layer is large. Particularly large depths tend to be associated with the many pools of standing water which occur at this site during the early summer (See Figure 5-10).

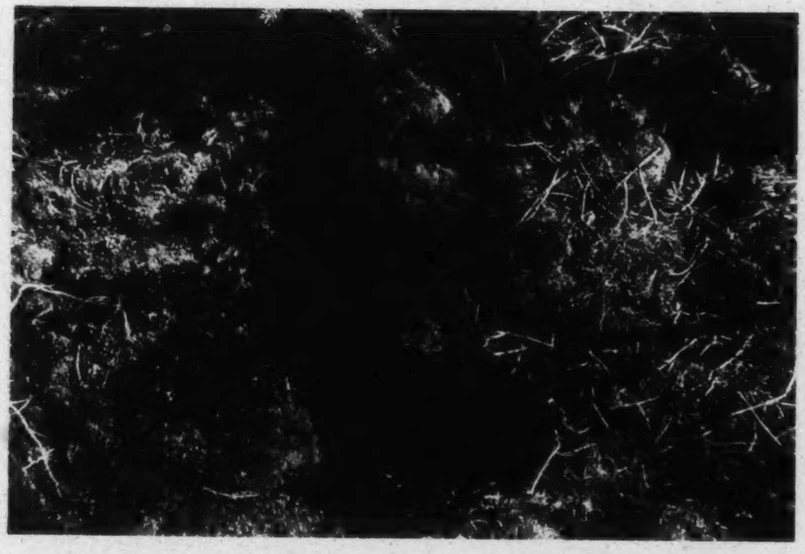
The Black Spruce Woodland bog, YT6, is found further along the Muskeg Creek trail from YT4, about 200 yards before the now disused gauging site for the sub-basin. The site forms part of a rim of woodland bordering a small lake which is slowly being filled in by the vog vegetation. A series of gently rounded ridge and pool areas occurs in the vicinity of the site which may be associated with long-term fluctuations in the level of the lake.

Two individual sites were investigated at YT6. The main site, YT6M, was located on one of the rounded ridges (Plate 4-8). This site seems to have been disturbed greatly during the caterpillar tractor activity associated with the ground temperature installations. The site is much less wooded than the majority of the bog. The moss cover was primarily Polytrichum commune and the organic layer was woody (Plate 4-9) extending to about seven feet in depth. At this depth bedrock occurs. Permafrost is found at this site, the average depth of the permafrost table at the end of the summer is two feet.

The sub-site investigated YT6S was selected for its

Plate 4-11: YT6S. Moss/lichen cover rolled back to show its relative thickness and looseness. Some *Cladonia chlorophaea* may be noted on the left of the photograph and *Polytrichum commune* on the right. 7

Plate 4-12: Pool 1, at corner of Muskeg Creek trail and Ryan Lake road. August 20, 1973. See also plates 5-2a and 5-2b. Photograph taken facing in a northerly direction.



greater representativeness with the surrounding spruce bog vegetation. It was well shaded by a relatively dense tree growth, (Plate 4-10) though located only about twenty yards from YT6M. The moss ground cover was quite loose and thick, (Plate 4-11). Polytrichum commune appeared to be the dominant moss but a lichen, Cladonia chlorophaea was locally abundant. The microrelief consisted of pronounced hummocks about one foot in height. Permafrost was present at a shallower depth than was the case at YT6M.

BEDROCK-DOMINATED SURFACES

A major limitation of the present study in terms of the objective to determine evapotranspiration patterns in the basin is the lack of field data for bedrock-dominated surfaces and transpiration through trees and shrubs. In the case of the bedrock surfaces it was possible to make useful conclusions from a consideration of the relationship between the nature of precipitation events and surface retention capacities. Some experiments have also been conducted with sprinklers to obtain information on summer runoff patterns from bedrock surfaces. Although the results of the experiments have not been published, personal communication with the researcher involved (A.G. Landals) helped to partly substantiate the general conclusions made. In the case of the significance of transpiration from the higher order vascular plants resort had to be made to the literature for indications of likely patterns.

BACKGROUND CLIMATIC DATA

Perspective on the conditions prevailing during the present study period in relation to previous periods or average conditions may be obtained from the background climatic data presented. These data were obtained primarily from the meteorological station operated by the Atmospheric Environment Service at Yellowknife Airport. Interpretation of results presented later indicate that temperature conditions at the sites investigated differ somewhat from Airport temperatures, and also that spatial variations in precipitation occur between sites. Although the discrepancies are not considered significant for the present study a mesoscale study of the climate of the region would be useful before attempting to extend the results of this study over a wider region. The influence of Great Slave Lake on the mesoclimate must be substantial. The lake may be expected to moderate the temperature of areas close to the lake. This may tend to dampen summer thunderstorm activity in the region. It would be very useful to delimit the areas affected by these possible influences. For example, Great Slave Lake appears to be the prime cause of the frequently substantial diurnal range in wind speed encountered in the study area during many of the summer days. The relatively high daytime speeds, according to the literature (Bliss, 1960) could be of major importance in limiting transpiration from trees and shrubs despite the existence of radiation conditions strongly favouring high rates of evapotranspiration. Presumably the lake's effect on wind speed diminishes further inland and the point where it dies out could be of significance in evapotranspiration studies. The meso-climate would appear to be a fruitful area for future research and a theoretical approach may be sufficient to produce worthwhile results if it is considered impractical to install numerous automatic weather recorders throughout the area.

RUNOFF

Daily discharge records for Baker Creek are available for the major, spring and summer, runoff period in each year since 1968 and these were obtained from the annual publications of the Water Survey of Canada, Environment Canada. The station name and number is Baker Creek Main Stem, 07SB009. The drainage area according to the published information is 55.7 square miles. For comparative purposes daily discharge records were similarly obtained where available for the Snare, Yellowknife and Lockhart rivers since 1965. All records were considered on a calendar year basis and monthly discharge was converted from acre-feet to inches of runoff averaged over the drainage area.

The quality of final records for Baker Creek has been described as generally good. Although it is believed that during the odd period daily means of estimated discharge may be slightly off, monthly means of good accuracy are expected (Wilson, Personal Communication, November 21, 1972). According to Landals (1970) however: "The natural control for the recording guage on Baker Creek is completely blocked by winter icing for most of the snowmelt runoff period.

Meaningful streamflow records, are unobtainable under such conditions" (p 9). Interpretation of the data indicated that icing development has only been favoured in the two years (1968; 1969) when Landals had the opportunity to observe spring snowmelt conditions. The hydrographs for these years contained a good length of rising limb and a peak encouraging the belief that gap periods could be estimated confidently. In addition icing may be anticipated to affect little more than locally derived snowmelt runoff rather than that derived from upper parts of the basin. The period of influence may be only a relatively few days in some years during rapid break-up or several weeks during extended break-up conditions. It is probable that gauge blockage problems associated with icing appear serious temporarily in some years. In exceptional years there may be justification for adding up to ten percent to the measured discharge. In relatively open years the upward adjustment may be as little as one or two percent (Laycock, Personal Communication, August 1973). These adjustments would include the winter flow that may not be measured by the gauge, but which is manifested by the icing development.

Data for the Yellowknife and Snare Rivers ceased to be published by the Water Survey from 1970. The reason given by a Water Survey official is one of doubts concerning the accuracy resulting from the manner of measurement (by power station personnel) and also the accuracy of the turbine ratings which are used to gauge a portion of the streamflow at each station. The status of the results from these stations prior to 1970 (and since) is therefore not clear. However in the absence of definite proof to the contrary the data have been accepted as reasonably accurate. Although the data are not now being published by the Water Survey it is still being collected by the companies. (Northern Canada Power Commission for Snare; Cominco Mines for Yellowknife River). Copies of the data for the Yellowknife River for 1970 and 1972, and for Snare River for 1971 were obtained through correspondence with the companies. It may be noted that for a few months in certain years discharge from the Lockhart River had to be estimated. This was accomplished with reasonable confidence since this river's regime pattern is fairly stable from year to year.

Another source of limitations pertaining to the streamflow data for Baker Creek is related to the practice of pulling the station's stilling well out in the fall and replacing it again in the spring. As a result of this practice it is impossible to determine the time when snowmelt runoff commences and when fall runoff ceases. In one year, 1971, the peak runoff had apparently occurred before the station was put into operation. Such difficulties had to be overcome by projection of the available trend to the likely date of commencement and termination of flow, which in many cases involved speculative estimation guided by a knowledge of concurrent temperature precipitation conditions and the patterns in other years. Snowmelt runoff is the most significant component of annual runoff. Except for 1971 this portion of the hydrographs could be estimated with a fair degree of confidence since the apparent peak was recorded and the rising limb gradient was well established. The fall runoff was generally very low except for 1970 and particularly 1969. Projections were made with reasonable confidence however for such periods. Where doubts existed projections tended to favour slightly high rather than slightly low values.

The focus on inches depth of runoff directs attention to the importance of the drainage basin area estimate. The difficulty of accurately distinguishing drainage divides in such remote areas of glacially deranged drainage cannot be overstressed. It is possible that the published values for Baker Creek could be inaccurate. Little can be done here to evaluate this possible error, or correct for it, except to make it known that there is the possibility of a slightly erroneous area value. To allow this study to proceed the drainage areas were used as published in all cases.*

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* The published area estimate was made by researchers working in the basin, associated with the Dept. of Geography, University of Alberta. Aerial photography and ten-foot contour interval topographical maps were employed initially. One of the major problem areas, to the southwest and west of Baker Lake and Martin Lake was checked out on foot by Dr. A.H. Laycock and A.G. Landals. The estimates may be out by a probable maximum of five percent. It is more likely to be much less with errors largely compensating one another (Laycock, Personal Communication, August 1973). The problem of making relevant contributing drainage area estimates for individual years is probably much more significant than any error in the published drainage area estimate. This is discussed more fully in the runoff section of Chapter Five.

LAKE EVAPORATION

Estimates of lake surface temperatures were required to enable use of the Richards and Webb (1971) equation for deriving estimates of lake evaporation. The climatological data necessary were obtained from Yellowknife Airport.

Several lakes or bays of varying size, shape and orientation were selected in the Yellowknife area for the purpose of investigating lake surface temperatures during the study period. Ease of accessibility by road was a major consideration in the selection process. Temperatures were measured by a standard mercury-in-bulb thermometer which was placed about one inch below the water surface. Fixed measurement sites were visited during the study period and data from these were supplemented by temperatures taken during a few canoe traverses of the lakes when assistance was available. The sites were visited at intervals of one to two weeks during the summer.

Supplementary data on the seasonal-depth variations in Grace, Kam and Frame Lakes, just south of Yellowknife, were obtained for the same period from researchers at the Freshwater Institute, Yellowknife. The measurements were made at the deepest points in each lake. These were at 18 m., 12.5 m., and 6 m. for Grace, Kam and Frame respectively. The mean depth of each lake is probably much less than these values. Temperatures were taken from thermistor probes, generally at intervals of one metre, about every two weeks.

The survey of lake surface temperatures was not rigorously conducted owing to the lack of continuously recording instrumentation. Also the need to attend to other areas of research prevented completely adequate attention being given to the survey. It must therefore be considered very preliminary in nature.

A particular limitation of the present work is the inadequate account given to considerable short period natural variation in lake surface temperatures which may be noted in the literature. Monthly means may be very unrepresentative for shorter periods and their use in equations could generate inaccurate estimations of monthly evaporation. Another limitation relates to the representativeness of Yellowknife Airport data on wind speed and vapour pressure for the lakes in the region. The latter area is also an important one for future research. For the present however the example of the Northern Ontario study (Richards and Webb, 1971) serves to indicate the utility of such coarse sample studies as an alternative means of deriving estimates of lake evaporation for the study area.

POOL EVAPORATION

A program of observation of the variations in water level of several intermittent pools in the study area was begun in early June during a period when the need to repair some instrumentation curtailed other research plans. It was assumed that for each pool seepage losses would be low or negligible and that any subsurface inflow would be balanced in large part by subsurface outflow. Since the pools had no surface channel outlet it was hypothesised that changes in level would provide relatively direct estimates of evaporation results.

A stake was driven into the bottom of the pools and a ruler attached firmly by wire. Readings were made generally on a daily basin though for a few days several readings were taken at various times of the day. The measuring sites were selected so that they could be easily visited in the course of other fieldwork.

The main pool monitored P1 (Figure 1-3) was located near the junction of the Muskeg Creek trail and the Ryan Lake road. It is a large saucer-like depression, approximately 8000 square feet in area and about eight feet deep at its deepest ppint. It appears to have been caused by excavation for garden soil material several years ago. Permafrost conditions exist in the areas bounding the pool but it was not possible to determine if such conditions also existed below the pool. Early in the season probing revealed frozen ground under the immediate edge of the pool to be at twice the depth found for the mainly moss covered surrounding area. It is likely that this discrepancy increased in magnitude in the course of the summer. General observations in previous years indicate that the pool almost dries out completely by the end of June (R. Steiner, Personal Communication, May 28th, 1973). It is filled up each spring by local runoff, mainly from an adjacent muskeg (Plate 4-12).

Four other pools were monitored at the sedge meadow, YT4. P2 is a rectangular, artificially formed pool, about 10 inches deep at its deepest point, and approximately 70 square feet in area at its maximum extent. Disturbance has revealed a very black organic soil floor. Permafrost occurs at all the sedge meadow sites though the active layer was always significantly deeper under the pools compared with the drier adjacent areas. P3, P4, P5 at YT4 are relatively similar being quite shallow, (generally no more than 10 inches deep at their deepest), and very small (20-50 square feet). Sedge grows in the pools and larvae float about the surface.

Although monitoring of the pools was not considered in the initial research design and hence does not fit in very well with other research, the program did produce interesting results in terms of possible advective effects and soil moisture gradient patterns. It would seem to be a good area for future research. A battery of different pool sites could be selected and viewed in terms of a black box. Each site would possess certain possible influences on water loss rates such as special surface, size, depth, permafrost characteristics. Response to a known storm event could be monitored to determine the relative levels of importance of the influences and to gain information on the local evaporation climate.

NET RADIATION

Net radiation flux provides a measure of the energy available at the ground surface prior to its partitioning to ground heat, sensible heat and evaporative heat fluxes. (Photosynthesis utilises a negligible portion of net flux. During the spring period a small but seasonally significant proportion of the net radiation is required for the melting of snow and ice) See Appendix C for information on the energy balance equation and the Bowen Ratio.

Although estimates of net radiation are available for Yellowknife Airport, (Wendland 1967, Hay 1970), it was considered desirable to attempt to obtain measurements of net radiation for certain specific surface types in Baker Creek basin. Owing to instrument problems it was only possible to obtain a useful record for YT2.

Net radiation was monitored at the muskeg site by means of a Thornthwaite net radiation recording system (Model 602-DC. 12v.D.C. CWT-1162B). Weather-proof housing for the instrument was constructed and the sensor mounted on a support at a height of three feet above the ground surface. It was not possible to incorporate a levelling device into the construction; however the sensor was attached firmly to the support and levelled carefully by eye whenever necessary. The surface area sensed by the radiometer, (approximately a circle with a radius of nine feet), was selected to include parts of as many microsites within the muskeg as possible.

The recording system was powered by a 12 v. auto battery which had to be recharged approximately every four days. Logistic problems associated with this requirement contributed to a few breaks in an otherwise continuous record from June 6th - August 31st. The output trace was reduced by hand or by computer digitizing to obtain daily estimates of net radiation in calories per square centimeter.

The net radiation estimates obtained in this study are presented with important reservations. It is very difficult to measure net radiation accurately. According to Hay (1970, p 122), the World Meteorological Organisation Radiation Commission (1961) gives an accuracy estimate of ±15 percent + 10 cal. sq. cm. for daily sums of instrumentally observed net radiation values. At high latitude sites considerable measurement error can result owing to the poor cosine response of the instrument at low sun angles. As a result of this some beam radiation may be reflected by the polyethylene windows of the net radiometer rather than being recorded. Wendland (1967, p 3) has estimated that because of this factor net radiation would tend to be undermeasured at Yellowknife by 15 percent. Orvig (1961) and Davies (1963) have estimated an almost similar undermeasurement at Schefferville.

Apart from the inherent errors mentioned above, problems of condensation inside the polyethylene hemispheres were a particular source of error. Some condensation was observed to occur on several occasions. This normally affected only night-time readings if the sensor was purged with dry air in the early morning. However, it was not possible to visit the instrument every day and it is likely that there were occasions when late morning values were affected. Another potential source of error is the lack of absolutely precise horizontal levelling of the sensor.

In view of the many sources of potential error hourly or daily data are not considered as useful as monthly means of daily net radiation and it is the latter which receive emphasis in the analysis.

ACTIVE LAYER DEVELOPMENT

The development of the active layer at the permafrost sites was monitored periodically during the summer by probing with a steel rod to determine its depth.

During each measurement period 25 sample probes were made over a fixed area of about 500 square feet. A stratified random sampling procedure was followed whereby typical microsite variations within the fixed areas would be sensed. At the spruce bog woodland, YT6, the microsites were: a) dense woodland; b) shallow pools containing sedge vegetation; c) exposed site denuded of trees. At the sedge meadow, YT4, the microsites were: a) sedge vegetation with no associated pools; b) sedge vegetation in pool or depression areas; c) scattered ground birch.

SOIL MOISTURE

Soil moisture patterns were monitored at the evaporimeter sites, partly to provide general information on summer regimes and field capacity relationships, but also to serve as a check on the behaviour of the evaporimeters in relation to natural conditions.

At each site, YT2, YT4, YT6S, YT6M; a fixed homogeneous area of about 200 square feet was selected which included the battery of evaporimeters. Moisture samples were taken from within this area approximately weekly and consisted of about six to ten in number, representing the top five inches (approximately 12.5 cms.) of soil. The surface was disturbed as little as possible. The samples were placed in a plastic bag and weighed immediately using a beam balance. At the end of the field season laboratory determination of the volumetric moisture contents of the samples was carried out by oven drying at 105°C for 24 hours to obtain the dry weight. The difference between the wet weight and the dry weight was divided by the wet weight and expressed as a percentage.

During July and August occasional pits were dug to a depth of about 30 inches (75 cms.) or to the foot of the active layer at the sites where permafrost occurred. Moisture samples were taken at frequent depth intervals in an effort to obtain some information on soil moisture/depth relationships.

It was necessary for comparative and interpretation purposes to obtain estimates of the field capacity of the top layers of the soil and the surface for each site. A typical area of about ten square feet was thoroughly wetted then covered with a groundsheet which was in turn covered with aluminium foil to reduce evaporation. After about 36 to 48 hours four samples each were taken from four layers between the surface and approximately six inches (15 cms.) depth. (Munn, 1966, p 27-28).

EVAPORIMETER STUDIES

An attempt was made to determine evapotranspiration directly by means of batteries of small juice can evaporimeters. Such instruments were originally employed by Mather and Thornthwaite (1956; 1958) at Point Barrow to determine P.E. They consisted of ordinary used metal fruit juice cans, 10.6 cms. in diameter and 17.4 cms. high. Inverting a cut down plastic container in the bottom of these cans provided a small water collection basin. A small polyethylene tube was inserted so that excess water could be removed periodically.

To install the evaporimeters a cylindrical mass of peat and surface vegetation was carefully cored using a bottomless juice can with a jagged cutting edge. The depth of core taken was about 5 inches (12.5 cms.) The core was inserted into the evaporimeter and the evaporimeter was placed in the hole from which the core had been removed. The hole had to be deepened slightly in order that when the evaporimeter was in place the top of the can was flush with the surface. The peat/moss material was found easy to work with and it is felt that the soil profile was only very slightly disturbed. Because of the very moist conditions prevailing in the soil at the time of installation the collection basins were almost, although not completely filled with water. This practice was to partly accommodate high levels of infiltration associated with a heavy rainstorm. Such events were infrequent during the study period however and it appeared that most of the summer rain was easily absorbed by the surface peat and moss layers. When installed each can blended in well with the microenvironment of the site, minimising rim effects (See Plates 4-2, 4-5, 4-8, 4-10).

The evaporimeters were installed in batteries covering an area of about 50 square feet. At YT2, YT4 and YT6M ten evaporimeters comprised each bettery while eight cans were installed at YT6S. Statistical treatment of results obtained from the numerous installations at Point Barrow during the 1956 studies indicated that 8-10 cans "would be most satisfactory" (Mather and Thornthwaite, 1956, p 27).

The cans were weighed periodically by means of a beam balance and the average change in weight of the cans was calculated in grams. From the can diameter it was deduced that a change in weight of 85 grams corresponded to a moisture gain or loss of one cm. Precipitation was measured at each site with a standard rain guage set on the ground surface. Evaporation from the cans was calculated by summing the change in weight expressed in centimeters of water and the precipitation, (if any).

Obvious limitations of the small can evaporimeters relate to their small surface area and their shallow depth. The former limitation is obviated to a reasonable degree by a battery, therefore their shallow depth becomes probably the most important limitation since the cans cut off the supply of moisture from, (or to), deeper zones.

During studies at Point Barrow simultaneous evapotranspiration measurements were made from small juice can evaporimeters and also from a large 50 gallon oil drum evaporimeter at the same sites. At one site (Aerology - low centred polygons) the small cans recorded 12 percent less than the large oil drum during the 1957 season. At another site (Britton Manor - high centred polygons) the smaller cans recorded 20 percent less in 1957 and 23 percent less in 1958, by comparison with the large drum (Mather and Thornthwaite, 1958, p 92). It was concluded that the instruments were not very suitable for measuring daily water loss. This conclusion also applied, but less markedly, for weekly estimates of water loss.

The Point Barrow experiments measured P.E. involving regular addition of water to ensure the soil moisture was at field capacity. In the present study attempts were made to monitor only actual evapotranspiration. At the Britton Manor site in 1957 a sample of cans was left exposed to natural conditions rather than periodic artificial irrigation. In effect these cans measured A.E. as in the present study. This group recorded 30 percent less than the large deep oil drum, and 20 percent less than the group kept at field capacity. This discrepancy could have been greater if it had not been for the fact that the precipitation was fairly well-distributed through the season entailing only slight variations in moisture content between the two groups. It may be concluded from the Point Barrow work that relatively large percentage underestimates of evapotranspiration can be anticipated when shallow cans are used, particularly when precipitation is erratic in time distribution as is the case for the present study area. Also the studies suggest that the reliability of the small evaporimeters is greatest when they are used to measure evapotranspiration for periods greater than a week.

Brown (1961) used three types of evaporimeters at one of his sites during the Norman Wells studies. These consisted of half a 45 gallon oil drum, (estimated depth 18 inches, 45 cms.), six 15 cm. diameter cans (estimated depth 12 inches, 30 cms.), and six 10 cm. diameter cans (depth 7 inches, 17.4 cms.). Using the average value of the six-can groups it was found that the shallowest cans recorded 40 percent more evapotranspiration than the other two types which were very similar to one another. This pattern is inconsistent with that noted in the Point Barrow studies, and also the pattern which may be deduced for the present study. The inconsistency is probably attributable to deficiencies in the method of sprinkling water on the surface to maintain P.E. conditions (Brown, 1961, p 24). The subjective judgement involved in this process has also been critically remarked upon by Mather and Thornthwaite, (1958, p 91), Nebiker, (1957, p 53, p 56), Watts, Galloway and Grenier, (1960, p 14). It is also possible that the inconsistency is partly attributable to the fact that the daily addition of water actually changes the climate under test (Brown 1965, p 28) and generates potentially significant advective effects. Since no artifical watering was undertaken in the present study these sources of error were avoided but there remains the strong doubt concerning the cans' ability to simulate natural soil moisture flow conditions. It was found possible to estimate the adjustment required for this factor by utilising soil moisture regime results and changes in weight of the cans.

A limitation of relatively minor importance associated with the evaporimeter studies is the necessity to introduce measured precipitation into evapotranspiration computations. The undermeasurement of precipitation attributable to characteristics of the rainguages is possible owing to wetting effects and evaporation from the guage. According to Bavina (1967, p 348): "This correction does not exceed 10 percent and amounts mostly to 3-5 percent of monthly evaporation for the summer in the forested zone of the USSR." This limitation appears to be of negligible importance but caution is suggested by the significant microsite variations in precipitation distribution for the high centred polygons (Mather and Thornthwaite, (1956,p 27).

EQUILIBRIUM EVAPOTRANSPIRATION MODEL

A simple model for determining evaporation from high latitude upland sites has been proposed by Rouse and Stewart (1972). The model is a derivation of the equilibrium model presented by Slayter and McIlroy (1960). It is just a simplified version of Penman's P.E. equation and it uses only air temperature and net radiation data, (and ground heat flux if available). As originally presented this model was considered applicable only under conditions of a saturated atmosphere. However further developments have revealed that the model is applicable also in non-saturated conditions. Wilson and Rouse (1972) found that the equilibrium model performed well over a moderately dry surface exhibiting a relatively high resistance to evaporation. Rouse and Stewart (1972) observed that high latitude upland sites in Northern Canada have vegetation covers, (non-transpiring ground lichen, mixed ground lichen, widely spaced conifers and lower order vascular plants), which exhibit a diffusive resistance to the efficient movement of water vapour from the moisture source to the atmospheric sink. This observation prompted the hypothesis that the equilibrium model or some function of it might be applied to estimate the evaporation for resistant subarctic and tundra surfaces. Rouse and Stewart (1972) successfully tested this hypothesis for a high latitude lichen-dominated upland site in the Hudson Bay Lowlands and suggested that the model could be used to reliably estimate evaporation for any high latitude surface which exhibits a relatively large resistance to evaporation. The appeal of the model stems from the fact that the only inputs required are net radiation, soil heat flow and screen temperatures. The prediction model by which LE can be determined in cal. per sq. cm. per min. was given as:

$$LE = LE_M = (0.406 + 0.011T)\ (Rn-G) \text{ where}$$

T = temperature in ^{o}C; Rn = net radiation in cals. sq. cm. min; G = soil heat flux in cals. sq. cm. min.
(Rouse and Stewart, 1972, p 1069).

In the present study this equation was applied to some data obtained during 24 hour intensive study periods - three each at YT2, YT4 and YT6M - conducted during July and August 1972. For each day the period of positive net radiation was considered and this was broken down into three or four blocks of time. Net radiation was obtained at YT2 from the recording radiometer and at YT6M and YT4 from spot portable radiometer observations. The ground heat flux was assumed to be 15 percent of the net radiation for each time period. This assumption is based partly on the results of the literature review reported previously (Chapter Three), the active layer development studies reported later (Chapter Five), and a qualitative assessment of the energy required to warm the ground at the sites. It must be considered very tentative however. The average screen temperature for each time period was used.

The data employed in the present study incorporate some assumptions which limit the results to the status of crude estimates. However the data are considered sufficiently adequate to indicate whether the application of the Rouse and Stewart model is satisfactory or unsatisfactory in the present study area.

Application of the model is warranted only when the wet bulb depressions of the overlying air (D) and the evaporating surface (Do) are similar. This situation is not likely to occur either when the surface is wet (D>>Do) or when it is very dry (D<<Do). For a middle range of moderately dry surface conditions where D - Do it is expected that the equilibrium model estimates will closely approximate the A.E. rate (Wilson, 1971, p6). Bowen ratio values equivalent to 'moderately dry' conditions have been indicated to be in the range o.25 - 0.54, Wilson and Rouse (1972). Where there is reason to suspect conditions where the Bowen ratio is outside this range the accuracy of the model predictions must be doubted.

CHAPTER V

RESULTS & INTERPRETATION

INTRODUCTION: It is the author's purpose in this chapter to present the results of the fieldwork undertaken in the course of this study and to supply an interpretation of the results. Major synthesis of the results at the basin and site water balance scale is deferred until Chapter Six.

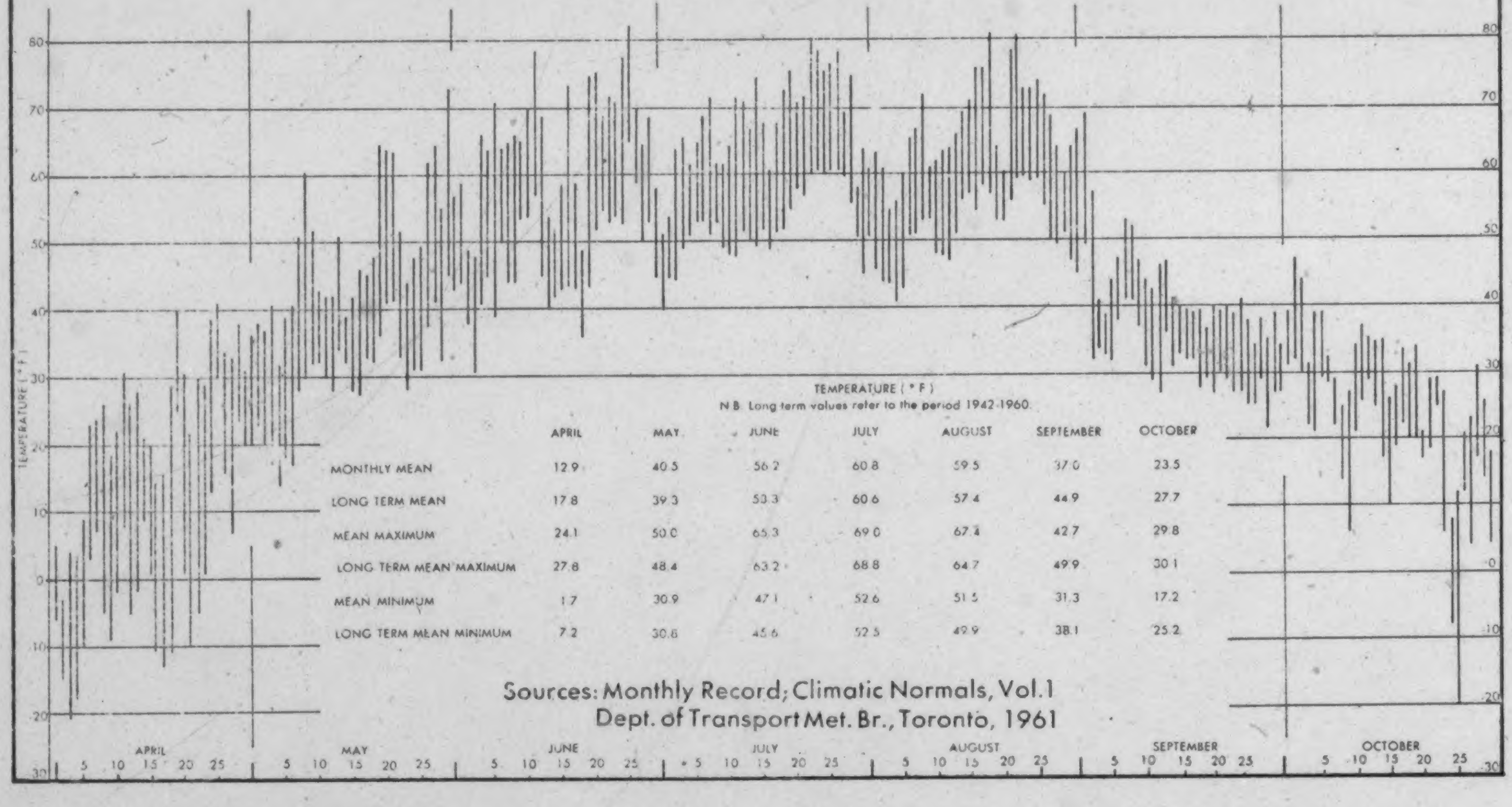
This chapter commences with the provision of background climatic data for the study season and the longer term trends. Particular emphasis is given to the period since 1967 because of the availability of runoff data for Baker Creek from about this time.

Following this survey, results pertaining to runoff, lake and pool evaporation, net radiation, active layer development, soil moisture, and the evaporimeter studies are presented in that order.

TEMPERATURE:

The daily maximum and minimum temperatures at Yellowknife Airport during the summer of 1972 are presented in Figure 5-1. After a relatively cool April, temperatures climbed quite rapidly in May and began levelling out about mid-June until winter arrived suddenly early in September. During the three main study months average temperatures were the rule in July while in June and August they were 2 to 3°F above normal. The representativeness of the 1972 summer temperatures by comparison with previous years may be appreciated by reference to Figure 5-2 where the monthly

FIGURE 5-1 DAILY MAXIMUM AND MINIMUM TEMPERATURES AND LONG TERM MONTHLY MEANS, YELLOWKNIFE AIRPORT. APRIL-OCTOBER 1972.



TEMPERATURE (° F)

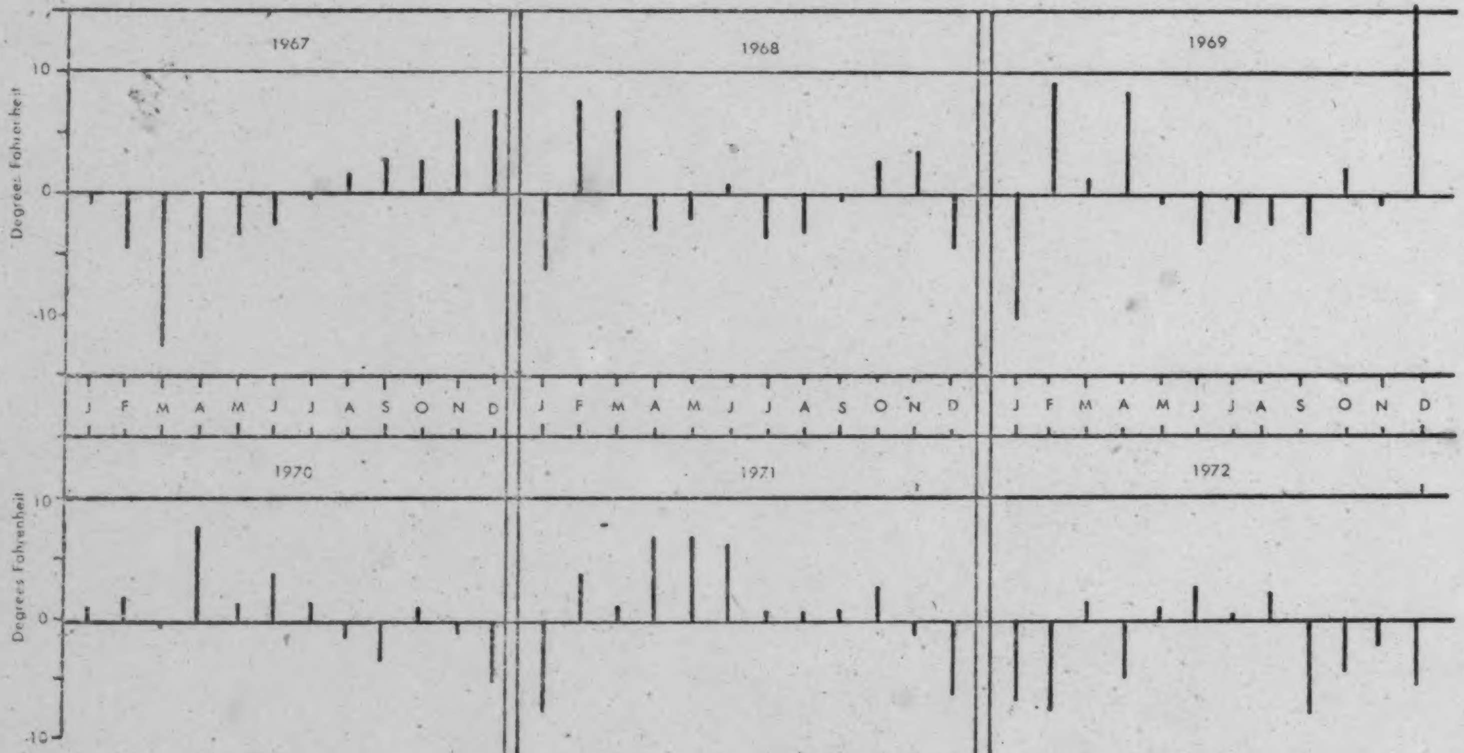
N.B. Long term values refer to the period 1942-1960.

	APRIL	MAY	JUNE	JULY	AUGUST	SEPTEMBER	OCTOBER
MONTHLY MEAN	12.9	40.5	56.2	60.8	59.5	37.0	23.5
LONG TERM MEAN	17.8	39.3	53.3	60.6	57.4	44.9	27.7
MEAN MAXIMUM	24.1	50.0	65.3	69.0	67.4	42.7	29.8
LONG TERM MEAN MAXIMUM	27.8	48.4	63.2	68.8	64.7	49.9	30.1
MEAN MINIMUM	1.7	30.9	47.1	52.6	51.5	31.3	17.2
LONG TERM MEAN MINIMUM	7.2	30.6	45.6	52.5	49.9	38.1	25.2

Sources: Monthly Record; Climatic Normals, Vol.1
Dept. of Transport Met. Br., Toronto, 1961

FIGURE 5-2 MONTHLY TEMPERATURE DIFFERENCE FROM NORMAL .YELLOWKNIFE AIRPORT, 1967- 1972

Source: Monthly Record



temperature difference from normal since 1967 is presented. Conditions in 1972 appear to have been relatively similar to past summers except for the slightly higher temperatures in June and August.

Areal variability of temperatures in the study area is not so marked as for precipitation. Mean temperatures at Yellowknife Hydro are normally about 2°F less than those at Yellowknife Airport during the summer months. (Table 5-1). Because of the low elevation of Yellowknife Hydro relative to the headwater region of the basin it is likely that mean temperatures during the summer in the latter area will be about 2-4°F on average below those in the southern portion of the basin represented by Yellowknife Airport. This difference could be great enough to introduce a significant contrast in evaporation levels between the northern and southern parts of the basin. Alternatively the 2°F difference between the Hydro and the Airport could possibly be due to measurement site factors. Although this is not thought to be likely it would have the effect of shallowing the temperature gradient over the basin and reducing temperature-induced evaporation contrasts.

On several days it was possible to compare hourly screen air temperatures at specific sites in the basin with those recorded at Yellowknife Airport. The results are presented in Figure 5-3. It may be observed that temperatures at the sedge meadow, spruce bog and muskeg sites tend to be lower than those at the Airport. The trend is particularly

TABLE 5-1. CLIMATIC NORMALS.
MEAN DAILY TEMPERATURE AND MEAN MONTHLY PRECIPITATION
YELLOWKNIFE AIRPORT AND YELLOWKNIFE HYDRO.

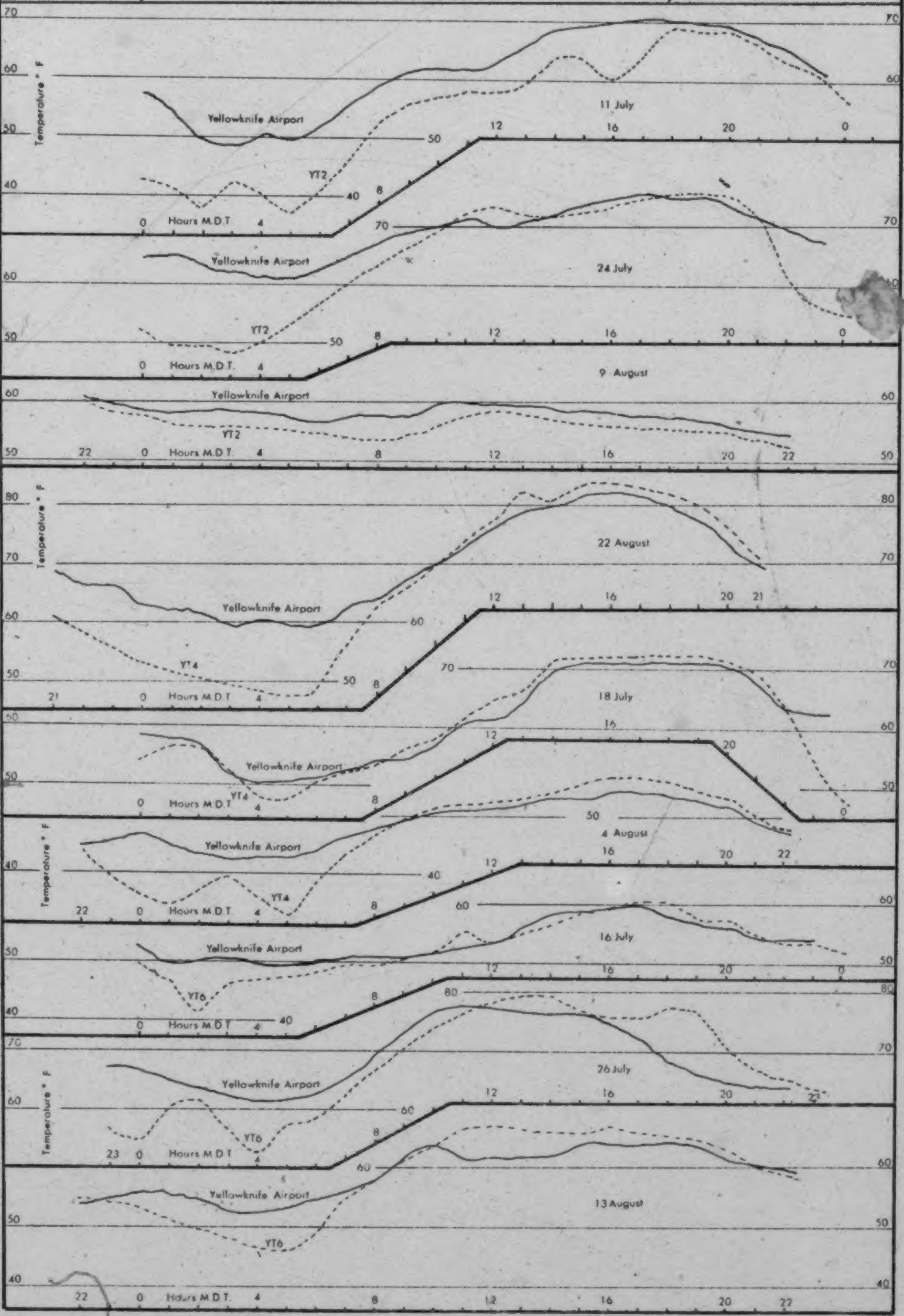
Source: Climatic Normals. vol.1.,vol.2., Canada, D.O.T., Met.Br., Toronto, 1961.

	Mean Daily Temperature (° F)		Mean Monthly Precipitation (inches)	
	Airport	Hydro	Airport	Hydro
J	-17.9	-17.1	0.51	0.50
F	-14.6	-15.5	0.49	0.45
M	-0.1	-1.0	0.51	0.48
A	17.8	17.1	0.39	0.43
M	39.3	37.8	0.66	0.61
J	53.3	51.1	0.64	0.71
J	60.6	58.4	1.42	1.65
A	57.4	55.0	1.39	1.66
S	44.9	43.4	1.16	1.24
O	29.7	28.7	1.10	0.97
N	7.2	6.6	0.92	0.87
D	-11.0	-11.8	0.81	0.85
Year	22.2	20.9	10.00	10.52

N.B. Normals are for the category where the period of record ranges from 10-24 years, during the period 1931-1960.

FIGURE 5-3. COMPARISON OF HOURLY TEMPERATURES AT BASIN SITES WITH YELLOWKNIFE AIRPORT DURING CERTAIN DAYS, SUMMER, 1972.

Source: Daily Surface Weather Record and field measurements by the Author



marked during the night and morning periods and is weakened or destroyed during the afternoon periods. It should be noted that temperatures were measured in standard Stevenson Screen installations at Yellowknife Airport and at the sample basin sites. All of the sites are effectively open in nature with regard to tree vegetation except for YT6 where the screen is located at the edge of the disturbed area. Here the effect of the spruce woodland may be expected to be significant in reducing air movement. This would encourage relatively high daytime air temperatures by comparison with less wooded sites. It may be concluded that the data presented in Figure 5-3 in general reflect the operation of this factor although the influence is not very marked.

Most of the days studied were dominated by sunny, clear or only partly cloudy conditions, breezy during the day and calm at night. Except for showery overcast periods (eg 9 August YT2), it appears from Figure 5-3 that during the day the Airport air temperatures tended to be slightly lower than the basin sites studied. This probably reflects the influence of a significantly higher albedo at the relatively unvegetated, sandy airport site by comparison with the others. Possibly the daytime discrepancy would be greater were it not for the stronger effect of evaporative cooling at the much moister vegetated basin sites. Limited emphasis should be placed on the latter possibility however in the absence of surface temperature comparative data.

The marked difference in nightime air temperatures between the airport and basin sites, present in most of the cases studied, may be due in large part to the most of the cases studied, may be due in large part to the effects of cold air drainage. The airport site is the only one not lying along a significant slope or near the trough of a depression. It is difficult to estimate the exact contribution of negative ground heat flux to this pattern in the absence of detailed measurements. Ground frost was observed on one occasion (4 August YT4) and quite heavy dew formation was noted on most remaining days with one definite early morning mist (18 July YT4). Qualitatively therefore it would seem that at the basin sites the release of heat stored during the day is also significant (in relation to cold air drainage effects). This implies that relatively inefficient insulation properties are present at the surface moss and peat layers when the opposite might be expected because of the often dried out nature of these layers.

From the cases studied it appears that mean air temperatures at peatland sites will be several degrees lower at the basin depression sites than those recorded at Yellowknife Airport. This is consistent with cases reported in the literature (Brown, 1965; Williams, 1968; Bavina, 1967). It would be useful to extend this type of study to embrace free water and bedrock-dominated surfaces and to measure both air and surface temperatures.

PRECIPITATION:

The precipitation for Yellowknife Airport during the main study period is presented in Figure 5-4. No snow precipitation was observed during this period. Several lengthy dry periods may be noted between the rain events. This effect is accentuated when it is appreciated that on at least five occasions a single rain event occurred over midnight necessitating a splitting of the total between two record days. Rainfall was generally light and of relatively brief duration except for a few heavy showers and one major thunderstorm.

The precipitation totals for these three summer months of interest were very similar to the long term average for each month except in the case of August (Table 5-1).

The summer as a whole however was not very similar to other summers during the previous five years (Figure 5-5). In particular it lacked the abnormally high late summer and fall precipitation amounts which characterised the previous five years.

Some authorities have indicated the probable need for substantial upward adjustment of the precipitation measured at stations in Northern Canada. The degree to which adjustment is necessary has been determined with reasonable confidence at only one location, (Knob Lake; Findlay 1969), where regular and extensive snow surveys have permitted adjustment of precipitation measured at the standard meteorological station. For the present study area, in

FIGURE 5-4. DAILY PRECIPITATION JUNE, JULY, AUGUST, 1972 YELLOWKNIFE AIRPORT

Source: Monthly Record

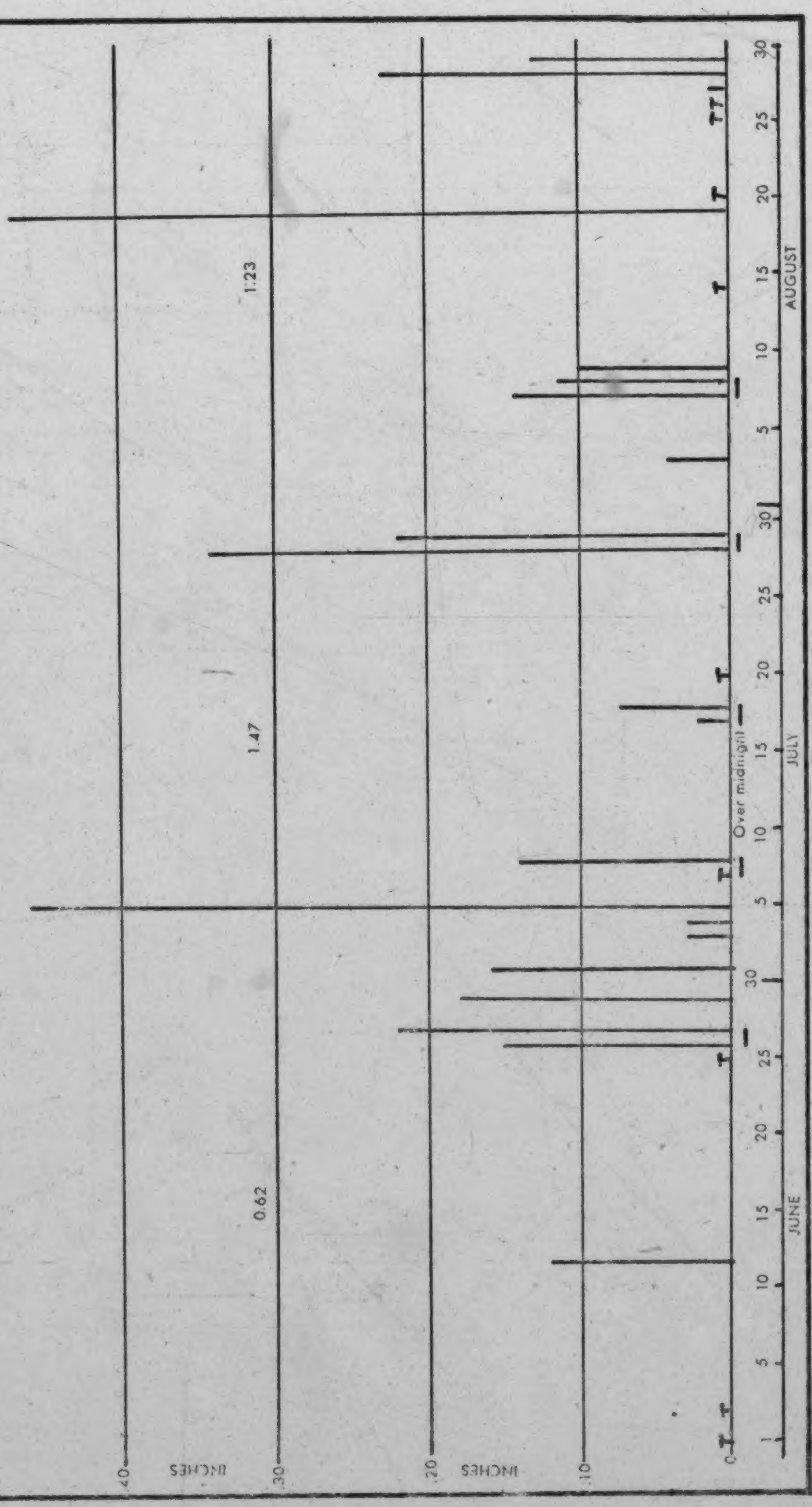
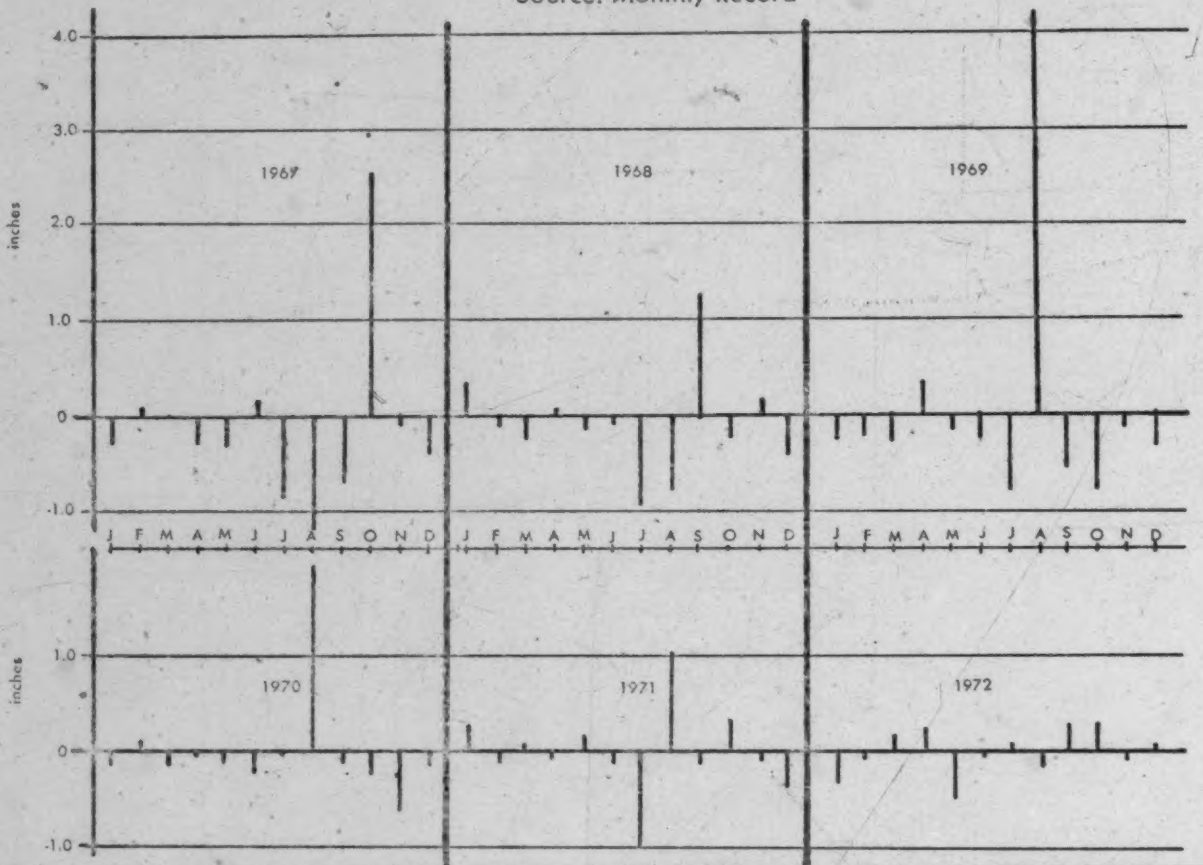


FIGURE 5-5. MONTHLY PRECIPITATION DIFFERENCE FROM NORMAL. 1967-1972.
YELLOWKNIFE AIRPORT

Source: Monthly Record



1967 Kakela (1969) made an extensive snow survey which indicated that Yellowknife Airport records undermeasured actual snow precipitation by 13 percent. Kakela thought this to be the direction of the adjustment but not necessarily its relative magnitude, which he inferred may be higher on average because of the influence of heavy snow precipitation years. In his calculations he used adjustment factors of +50 percent for snow precipitation and +15 percent for rain precipitation, following reports of Northern U.S.S.R. experience. Results from the present research indicate that these adjustments may well be large overestimates. In Table 5-2 is presented the unadjusted total precipitation for the years 1968-72 and adjusted totals for two combinations of adjustment factors. The present research suggested the adjustment necessary probably lies somewhere between the unadjusted total and the first combination.

Precipitation amounts measured from standard rain gauges at specific sites in the study region are compared with Yellowknife Airport data in Table 5-3. Significant differences may be noted between the areas in each month but no pattern is evident over the whole period. Similar areal variation has been noted for the region by Taylor

TABLE 5-2

Total Annual Precipitation
Unadjusted and sample adjusted totals
Yellowknife Airport 1968 - 1972
(inches)

	1968	1969	1970	1971	1972	Mean
Recorded (no adjustments)	8.80	11.31	10.03	9.94	9.92	10.00
+ 10 percent Rain + 30 percent Snow	10.53	13.13	11.67	11.63	11.92	11.78
+ 15 percent Rain + 50 percent Snow	11.60	14.18	12.63	12.82	13.23	12.89

Source: Monthly Record. Calculations by author.

TABLE 5-3. SPATIAL VARIATION OF PRECIPITATION IN THE STUDY REGION.

June, July, August, 1972.

Sources: Monthly Record and field observations.

(inches)

			Pocket Lake		Muskeg Creek			
	YK.'A'	YT2	YT3 *	YT4	YT5 *	YT6	Hydro.	Discovery **
June	0.62	0.66	0.65	0.75	0.77	0.73	0.62	> 1.79
July	1.47	1.11	1.08	1.46	1.35	1.43	0.81	0.82
August (to August 27th.)	0.87	0.81	0.96	0.67	0.66	0.64	0.56	0.95

N.B. * YT3 and YT5 are rock sites associated with the ground temperature studies.

** Discovery is located 50 miles north of Yellowknife. A rain guage was operated there from June 22 to August 27.

(1971) in a comparative study of precipitation for Yellowknife Airport, Yellowknife Hydro and Discovery. The study included a comparison between Yellowknife Airport and Discovery for precipitation in July and August 1971. Total precipitation at Yellowknife was 25 percent greater than at Discovery. Daily patterns exhibited large differences which indicated not only the effects of local thunderstorm activity but also a considerable variation in rainy periods, when greater uniformity of precipitation might have been assumed over a distance of only 50 miles. It was concluded that owing to the significant variation between stations no one station could be taken as being any more representative of patterns in the region that others.

Yellowknife Hydro appears to be well located to serve as an indicator of the climate experienced by the headwater areas of Baker Creek basin. However, caution should be exercised in this regard. Apart from the evidence presented above for considerable areal variation in precipitation it should be noted that the elevation of Yellowknife Hydro (523 feet a.s.l.) is less than that at Yellowknife Airport (682 feet a.s.l.). Elevation in the headwater region of the basin ranges from about 700 to 850 feet.

It should be borne in mind that at Hydro, snow precipitation is measured with a ruler to determine depth and the water equivalent is obtained by dividing the depth by 10. At the Airport snow is measured by means of a Nipher shielded gauge, and is melted down to obtain the correct water equivalent. Winter water equivalents of snow precipitation calculated from records for Yellowknife Airport. for the winters 1966/7 to 1971/2 are presented in Table 5-4. They were calculated from the published data simply by dividing the precipitation total given for the month by the depth of snow in inches. Assuming water equivalent patterns at the Airport are typical of the surrounding region the data indicate that the 0.10 equivalent used for the Hydro station is too high. It is suggested that a more appropriate conversion factor for Yellowknife Hydro snow precipitation is approximately 0.075. This would have the effect of reducing annual hydro precipitation by about one inch to a mean level below that for the Airport. The elevation difference between the Hydro and the headwater area of the basin may compensate for this reduction to some extent, minimising contrasts in mean annual precipitation over the basin. It is of interest to note however the consistently greater precipitation recorded for the Hydro during the summer months by comparison with Yellowknife Airport. This may be a function of Hydro's valley-foot location possibly favouring the funnelling of storm activity over the station with less effect being felt in higher areas removed from the valley such as the upper Baker Creek basin. If this funnelling pattern is not present and the upper basin receives slightly greater summer precipitation, on increased elevation grounds, than Yellow-

TABLE 5-4

Monthly average water equivalent of snow precipitation calculated from Yellowknife Airport data. Winters 1966/7 to 1971/2.
(inches of water equivalent per inch depth of snow)

	66/7	67/8	68/9	69/70	70/71	71/2
October	.112	.083	.093	.100	.076	.083
November	.074	.071	.081	.075	.060	.050
December	.070	.072	.080	.065	.070	.083
January	.070	.082	.065	.061	.081	.062
February	.078	.065	.076	.057	.080	.075
March	.085	.082	.063	.051	.115	.070

Source: D.O.T. Monthly Record. Calculations by the author.
Refer to the text for the method of calculation.

knife Hydro, which in turn is greater than that for Yellowknife Airport, it is possible that slightly more runoff from the upper basin will be favoured relative to the lower basin. This is suggested by Kakela's good correlation between annual runoff and summer precipitation (Kakela 1973, p 175).

CLASS 'A' PAN:

The Class 'A' Pan data for Yellowknife Airport during the 1972 season is presented in Figure 5-6. The data are in the form of values corrected for heat transfer effects associated with the pan. Relatively high rates of daily evaporation occurred in June and July, (mean monthly rates are 0.20 and 0.22 inches per day respectively, falling off in August (mean of 0.14 inches per day), to 0.07 inches per day on average in September. It may be assumed that a significantly high rate of evaporation persisted during the latter half of May, perhaps averaging 0.15 inches per day. Figure 5-6 may be taken as an index of the relative values of potential evaporation throughout the study period. The validity of the magnitude of the rates of evaporation for environments other than that of the pan is not certain.

The representativeness of the data for the summer of 1972 may be guaged from Table 5-5. Estimates have been included for May and parts of June and September where the records are missing. The estimated values are in brackets. The summer totals appear to be very high when set beside some of the values published in the literature. It may be

FIGURE 5-6. CLASS 'A' PAN EVAPORATION, (CORRECTED), FOR YELLOWKNIFE AIRPORT.
SUMMER, 1972

Source: Monthly Record.

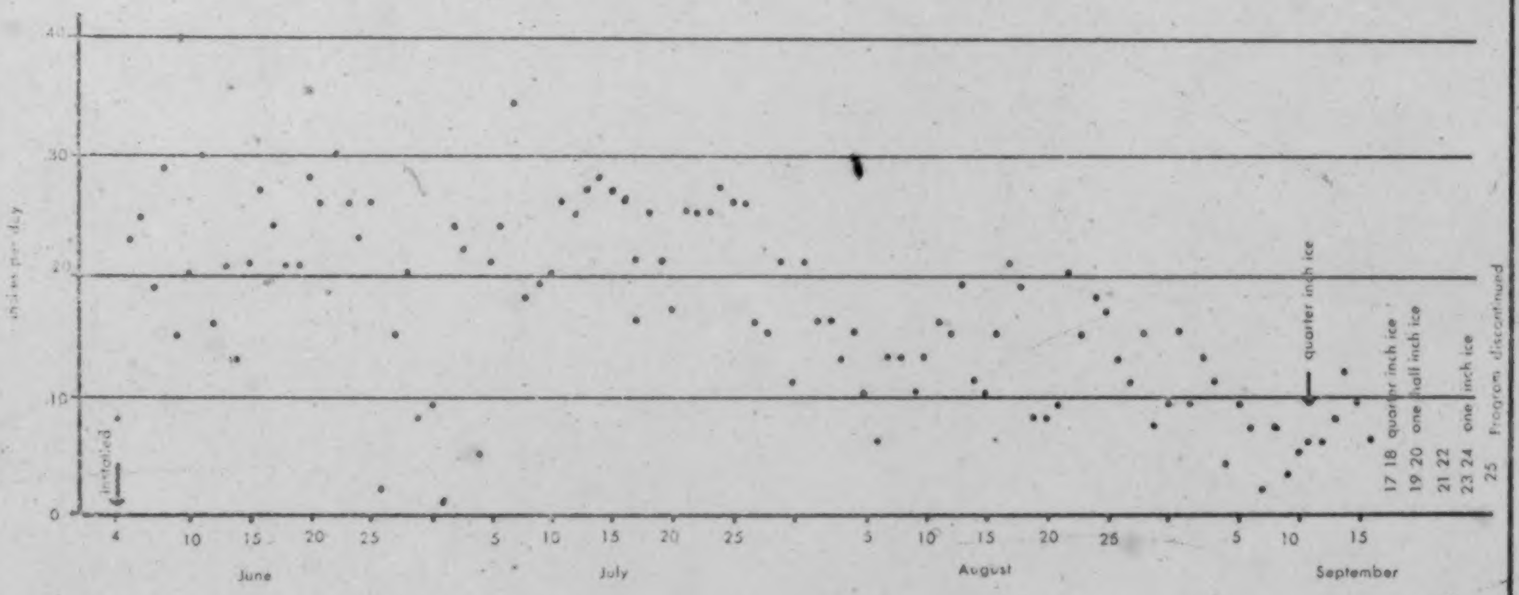


TABLE 5-5

Monthly Class A pan evaporation (corrected)
Yellowknife Airport 1966 - 1972
(inches)

	May	June	July	Aug	Sept.	Oct.	Total
1966	(2.00)	3.49 15/6 (2.50)	6.58	5.19	1.97	-- (0.30)	22.03
67	(2.00)	5.92	5.87	5.30	2.24	(0.30)	21.83
68	(2.00)	4.90 5/6 (0.75)	6.01	4.82	1.81	(0.20)	19.74
69	(2.00)	5.66 4/6 (0.60)	6.07	3.29	1.87	(0.20)	19.69
70	(2.00)	5.61 3/6 (0.40)	6.05	3.33	1.29		18.68
71	(2.00)	7.19	6.30	4.14	2.27	(0.50)	22.40
72	(2.00)	5.46 4/6 (0.60)	6.64	4.16	1.17 16/6		20.03
						Mean	20.63

Source: Monthly Record
Estimates by author.

noted that the total for 1972 is the closest to the 7 year average.

The pan evaporation records available for Yellowknife do not appear to have been very closely considered by Ferguson, O'Neill and Cork (1970) since the estimates provided by these workers are well below those recorded. Interpretation of results presented later indicates that the actual average for small lake evaporation in the study area probably lies midway between the averages extracted from the literature and from the pans.

RUNOFF:

The runoff regimes for Baker Creek during the period 1968-1972 are shown in Figure 5-7. Monthly runoff estimates were derived from the regimes and are presented in Table 5-6. The volume of streamflow has been expressed as depth of runoff in inches over the drainage basin. It has been pointed out earlier that owing to the potential error in estimating drainage basin area for such heavily glaciated terrain the values presented may be slightly inaccurate.

The table indicates large variations in runoff on an annual and monthly basis. In particular it may be noted that the runoff during 1972 was considerably lower than in previous years.

In order to interpret the above variations it is useful to note the results of research into drainage area as a hydrological factor in the Canadian Prairies (Spence; 1972; Stichling and Blackwell 1958). This research has singled

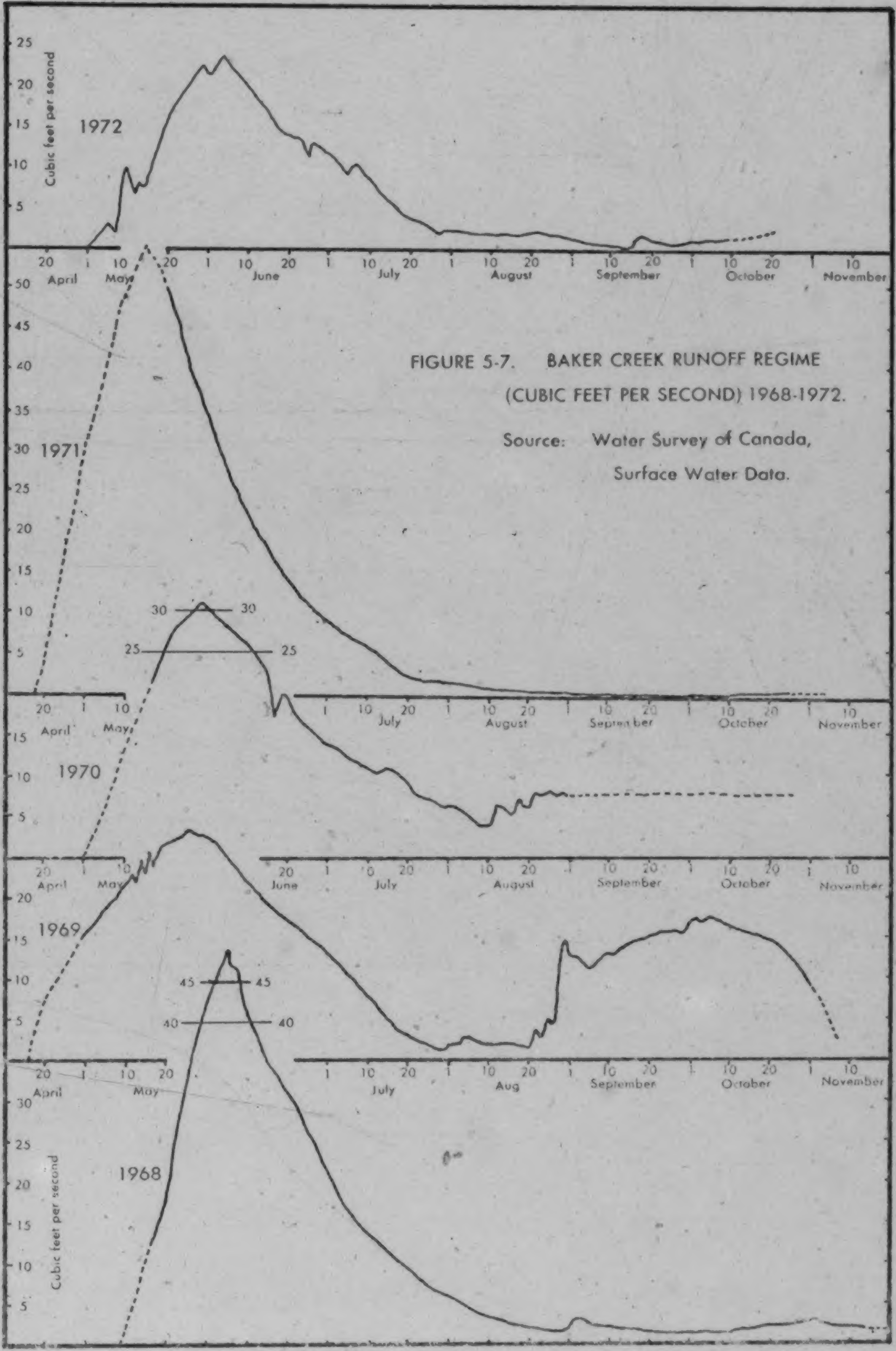


FIGURE 5-7. BAKER CREEK RUNOFF REGIME (CUBIC FEET PER SECOND) 1968-1972.

Source: Water Survey of Canada, Surface Water Data.

TABLE 5-6. MONTHLY RUNOFF: BAKER CREEK 1968-1972.

		April	May	June	July	August	September	October	November	TOTAL
1972										
	ac.-ft.		707	982	461	108	45			2302
	inches		0.24	0.33	0.16	0.04	0.02			0.79
1971										
	ac.-ft.	333	2998	1120	277	31	5	11		4775
	inches	0.11	1.01	0.38	0.09	0.01	0.002	0.004		1.61
1970										
	ac.-ft.		1076	1360	622	366	446	278		4148
	inches		0.36	0.46	0.21	0.12	0.15	0.09		1.39
1969										
	ac.-ft.	285	1490	1240	372	198	835	949	119	5448
	inches	0.09	0.50	0.42	0.13	0.07	0.28	0.32	0.04	1.84
1968	ac.-ft.		1094	2170	764	171	125	104	70	4498
	inches		0.37	0.73	0.26	0.06	0.04	0.04	0.02	1.52
									Mean	1.43 inches

Source: Water Survey of Canada; estimates by author out the depression storage characteristics of drainage areas as being of major significance in explaining runoff variations. Extensive glacial deposition, in combination with relatively low topographic relief, has created large areas of internal drainage into local depressions, many of which have no outlet except by evaporation, while others overflow and contribute to streamflow only in some years. Ideally a different drainage area to suit the climatic conditions of a particular period should be used.

Stichling and Blackwell (1958) have conveniently listed the factors which cause the boundaries of a watershed to fluctuate. They also note that these are analagous in many ways to the factors which contribute to the variability in runoff.

Where depression storage is appreciable it will effectively reduce the rate and the total volume of surface runoff. Soil and topography greatly affect infiltration which in turn influences the surface runoff and the rate at which depressional storage is filled. Precipitation and evaporation behave antithetically but they must be considered together as being the most significant variables affecting the limits of a drainage basin. The actual effect of precipitation will be determined by its seasonal distribution and must be considered in two parts, namely, that precipitation which produces the runoff and that precipitation which is antecedent. Evaporation is a factor which also affects antecedent conditions since high evaporation may minimise the effect of high antecedent precipitation and vice versa. Intensity of rainstorms and rate of warm-up in the spring influence the amount of infiltration and subsequently the volume of surface runoff which replenishes the depressional storage.

Consideration of the above factors greatly aids the interpretation of runoff variations for Baker Creek. Depressional storage is certainly appreciable. Glacial erosion and deposition have contributed to the formation of an intricate mosaic of depressions which are very loosely interconnected. It is probable that Baker Creek owes its status as a significant flowing water body to the exceptional interconnection properties afforded by the West Bay Fault which dominates the spine of the basin. A short distance away from this narrow corridor it is likely that many semi-permanent internal drainage basins dominate the drainage area.

It is probable that the low runoff for 1972 was in strong part due to the necessity of filling a large amount of depression storage which had been depleted by low precipitation or high evaporation in the previous period. It may be noted that the Class 'A' pan record for 1971 was the highest since measurements began, contrasting markedly with the record for the previous three years (Table 5-5). Since annual precipitation during the past three years has been very close to the average, high evaporation in 1971 was probably primarily responsible for depleting the depression storage. Other insights into possible explanations for the runoff variations may be deduced from Figures 5-2 and 5-5. From these it may be noted that in the summer of 1971 August was very wet but this seems to have been balanced by the dry July which preceded it since no significant late summer or fall runoff occurred. It may be significant however that the fall of 1971 was relatively warm and no particularly wet conditions preceded freeze-up. Such conditions would have ensured that depression storage would be well below capacity when 1972 breakup occurred. In addition it may be noted that breakup was relatively late in 1972, April being particularly cold, and there seems to be a direct relationship between high levels of runoff and early breakup eg. 1969, 1971. (1971, having a relatively warm April is an exception in this regard but the low level of snowmelt runoff appears attributable to relatively low winter precipitation, especially during late 1970).

The need to fill particularly depleted depression storage only seems to have arisen in 1972. The summer of 1968, except for September, was very dry suggesting that depression storage would have been at relatively low levels. However temperatures were also low during the summer of 1968. This would have reduced evaporation demands. In addition the wet September of 1968 would have topped up storage levels somewhat. In view of the relatively low precipitation which occurred in the winters of 1967/8 and 1968/9 the high snowmelt runoff in 1968 and 1969 is surprising. No singularly good reason appears to exist to explain this pattern except for the possible influence of icings. Icing is definitely known to have affected the valley of Baker Creek near, and including, the guage area in 1968 and 1969 (Landals, 1970, p 41). Landals (1970, p 41) has remarked on icing acting as a reservoir for retention storage: "Runoff is reduced throughout the winter period as it accumulates in the form of ice. Water stored in this manner is released in the spring, thereby substantially increasing runoff for this period." This would appear to contribute a large part of the explanation for the high levels of snowmelt runoff during the spring of 1968 and 1969.

The months of October and November for 1967 and 1968 were both characterised by above average temperatures. In no year since has this pattern been observed for this period, although the temperature for the month of October was slightly above normal in 1969, 1970 and 1971. October 1967 and September 1968 were abnormally wet months and temperature conditions favoured conversion of this precipitation into the icings. In no year since has this combination of circumstances been so prominent. There is an indication that conditions favouring slight icing occurred in the fall of 1971, but a field survey in the spring of 1972 revealed that icing was not significant.

The above association of surprisingly high snowmelt runoff periods with definite reports of significant icing development may contain a certain element of overemphasis. Winter icing development is also a very good indicator of high detention storage levels. The high levels of snowmelt runoff in the spring is partly attributable to melting of the icings but it is possibly more due to the fact that there is little unused detention storage capacity available to take up part of the local meltwater flow.

The high snowmelt runoff for 1970 appears to be attributable to the exceedingly high precipitation in August 1969. Only a small proportion of this record-breaking precipitation appears to have contributed to runoff in the fall of 1969. Much of it appears to have gone to fill depression storage to its total capacity. Precipitation during the winter of 1969/70 was below average and the high runoff flows during the spring and summer of 1970 may be explained in part at least by reference to the August 1969 precipitation. The runoff pattern associated with this precipitation event indicates that as a result of it depression storage was filled to overflow levels for much of the basin. It also indicates the existence of a considerable lag in time, approximately ten months, until the last effect of the event on runoff can be observed.

The very high snowmelt runoff in 1971 appears to be attributable mainly to a particularly rapid and intensive warming during the spring period. Another favourable factor would be the high probability that depression storage was relatively well filled as a result of the high precipitation during the months of August 1969 and 1970. Precipitation during the 1970/71 winter was about average on balance indicating the importance of the above factors.

Some indication of the changes in depression storage in the study area since 1966 is provided by reference to Pocket Lake lake level records (Station No: 07SB008). The approximate lake levels for the month of December were extrapolated from Figure 5-8. The results are shown in Table 5-7.

TABLE 5-7

Pocket Lake. Level changes Dec. 1966 - Dec. 72.

December	'66	'67	'68	'69	'70	'71	'72
Level (feet)	96.1	96.7	96.8	97.4	97.3	96.8	96.9
Change over year (feet)	---	+0.6	+0.1	+0.6	-0.1	-0.5	+0.1

Source: Water survey of Canada; and estimations by the author.

Regarding Table 5-7, the 1966 estimate is low in part because of the heavy drainage of the lake during construction of the weir and the lake not rising to overflow levels after the construction period. Construction of a new weir in 1967 did not affect levels significantly (Laycock, Personal Communication, August 1973). In interpretation of the table therefore emphasis should be placed on the direction and magnitude of the annual change in level, rather than on the absolute level values. Significant increases in Pocket Lake basin storage occurred from December 1966 to December 1969. After this it fell, particularly during 1971. The latest estimate indicates that some storage buildup occurred in 1972. This trend corresponds well with the pattern deduced from runoff variations.

As a result of the lake temperature survey at Prosperous Lake information was obtained on recent lake level changes from summer cottage owners on the lake side. Their information backed by markings on boat marina supports indicated that between the early summer of 1971 and the spring of 1972 the level of the lake fell by three feet. By the end of the summer of 1972 it had risen by about one foot (Plate 5-1). These respondents reported similar changes for Prelude Lake over the same period. Observations by the author at a number of smaller lakes in the study area consistently indicated a prominent wash line on rock surfaces abutting into the lake at about two feet above the existing lake level. These observations suggest further support for the storage change patterns identified above.

It is possible that important long, as well as short, wavelength cycles of change occur in depression storage in the basin. The signs of vegetation rims at the spruce woodland bog site have already been mentioned. In his study Brandon noted that: "Indications of high water loss by evaporation in summer can be seen north from Yellowknife where small closed lakes show varied vegetation rims around their edges " (Brandon, 1965, p 7). These observations indicate that the hydrological record of the basin could be successfully extended back in time by biogeographical studies, incorporating

Plate 5-1: Water level changes at Prosperous Lake. The person is pointing to the probable level of the lake during the early summer of 1971. After falling three feet by the spring of 1972 it had risen by one foot to the level at the date when the photograph was taken. (August 20, 1972).

peat analysis and dating, of several lake borders in the area.

The runoff for Baker Creek over the past five years has been compared with similar records for rivers in the broader region also guaged by the Water Survey of Canada (Table 5-8). It should be remembered that the values for Baker Creek contain an element of estimation for runoff during periods prior to installing the guage in spring and subsequent to taking out of the guage in the fall. The degree of estimation in the totals may be appreciated by reference to Figure 5-7 where dashed lines indicate the estimated portion of the regime. It was assumed that flow stopped during the main winter months. It was also assumed that early and late period runoff could be represented by an approximately linear projection of available records. Generally however the gap period estimations account for a relatively small proportion of the annual total and associated errors have been considered as not serious, except possibly in the case of the 1971 total when the peak runoff value and date had to be estimated. It is clear that Baker Creek has the lowest yield of the four rivers under consideration. In the comparison all the rivers have been treated as a lumped system whereby the values given represent an integration of runoff effects over the whole basin. This form of treatment is relatively satisfactory for very small basins such as Baker Creek. However, when considering such large drainage basins as the others in this comparison it is important to appreciate that large variations in runoff may occur within the basins. This point particularly applies in the

TABLE 5-8

Runoff for Selected Rivers in the Central Mackenzie District, N.W.T. 1965 - 1972

River:	Baker Creek	Yellowknife R.	Snare R.	Lockhart R.
Drainage Area (sq. mls.)	55.7	4350	5880	10,300
Calendar Years		(Total Inches Runoff)		
1965	--	2.56	3.13	5.60 *
66	--	2.58	3.43	5.60 *
67	--	3.06	4.09	5.46
68	1.52	4.52	4.46	6.30
69	1.84	4.12	4.46	6.00
70	1.39	4.91	M	6.00
71	1.61	M	4.52	6.30
72	0.79	4.08	M	5.96 *

M = Missing data   * Part of year estimated.

Sources: Water Survey of Canada, Personal Correspondence.
Calculations by the author.

case of the Snare, Yellowknife and Lockhart rivers because of their general location across a zone where bioclimatic gradients are steep (Hare and Ritchie, 1972).

It is possible that precipitation is greater in the relatively high headwater regions of the latter three large rivers. However very few measurements exist and the precipitation trend decreases slowly in a northeasterly direction indicating that increased precipitation may not be a very significant contributing factor to the increased yields by comparison with the southerly located Baker Creek. Caution is recommended in this regard however in the light of a recent paper by Findlay and McKay, (1972). Study of recently available snow course data, when mapped, indicated a wedge of surprisingly high water equivalent extending east and west through central Keewatin and Mackenzie districts. They add: "What has been traditionally described as the polar desert becomes much less arid when the mass of snow is considered instead of the depth" (Findlay and McKay, 1972, p 141).

What seems more certain is that the precipitation that does fall in the headwater regions is more effective because of the relatively cold temperatures which significantly reduce free water evaporation. The cold soils associated with the more widespread permafrost, and the much reduced tree cover can also be expected to reduce evapotranspiration. The Lockhart River drains a particularly large area in the region north of the treeline and within the zone of continuous permafrost. It may be noted that the yield for this river is significantly larger than for either the Snare or the Yellowknife Rivers.

The yield patterns for Baker Creek probably most accurately represent only the extreme southern portions of the other river basins. It may be inferred that average yields in the upper portions of the Yellowknife and Snare river basins could well be in the range of six to seven inches, and possibly rise to about eight inches in the case of the upper Lockhart basin.

Kakela (1969) calculated the twenty-five year mean runoff of the Yellowknife River to be 3.03 inches. This was on a water year basis. It may be noted that the runoff from the Yellowknife River during the period 1968-1972 appears to have been significantly above average. If Baker Creek behaves in sympathy with the larger regional variations this would indicate that the available Baker Creek record is representative of a relatively high runoff period. However the mean of the annual precipitation totals for Yellowknife Airport from 1968-72 is very similar to the long term average suggesting that the recent above average trend in Yellowknife river runoff is due to upper basin variations in precipitation or precipitation effectiveness. This reasoning also indicates that the 1968-72 record for Baker Creek is a good indicator of average conditions rather than of a particularly wet or dry period.

LAKE EVAPORATION:

The survey of lake surface temperature patterns was undertaken to allow application of a semi-empirical equation to obtain estimates of evaporation. The coarseness of the sample in the time dimension and the natural variability of the parameter resulted in a relatively wide scatter of points when the temperatures were plotted for each lake over the summer period. Broad trends were revealed however and it was possible to extrapolate tentative mean monthly lake surface temperatures. Some check on the rather crude estimates provided by this procedure was possible in the case of Kam and Grace Lakes by utilising the temperature-depth profile data collected by the Freshwater Institute researchers. The latter data were believed to be capable of providing rather more accurate estimates of monthly means because of the information the profiles provided on the behaviour of the hypolimnion and the thermocline in each lake. An appreciation of this behaviour aids in distinguishing the monthly pattern from the influence of daily variations although a certain element of dubious extrapolation is retained. As a result of this check good agreement was obtained for July and August between the monthly means extrapolated from the thermometer plots and those from the depth-temperature profile data. Study of the profiles for June indicated the influences of differences in depth which had not been sufficiently allowed for in the first set of extrapolations. The latter process is based on an assumption of a broad direct relationship between lake area and lake depth and hence lake volume. Although Kam Lake has a much larger area than Grace Lake the depth information indicates that Grace Lake contains a greater volume of water. Grace Lake was seen to be slightly cooler than Kam Lake rather than the reverse having been the case. Some adjustment to the original extrapolated values was necessary in view of this finding and it is possible that depth factors may similarly operate at other lakes though no data exist which could act as a guide for adjustments. It may also be noted that the profile data for Frame Lake, which was not monitored in the present study were observed to agree well with the extrapolated values for the adjacent and apparently similar Stock Lake.

The results of the lake surface temperature studies are presented in Table 5-9. This table contains an estimate of the approximate surface area of each lake. It also includes the monthly evaporation computed from the equation used by Richards and Webb (1971, p 7). Inputs to the equation were estimated mean monthly lake surface temperatures and climatological data on wind speed and vapour pressure of the air for Yellowknife Airport meteorological station. Estimated totals of evaporation include two inches to cover evaporation during May and September. This estimate is probably too low for Grace, Walsh and Prosperous Lakes since late season evaporation will largely compensate for the low early season evaporation and reduce the disparity with seasonal evaporation totals for other lakes in the basin.

The level of lake temperatures during the three main

TABLE 5-9. EXTRAPOLATED MEANS OF MONTHLY LAKE SURFACE TEMPERATURE AND COMPUTED EVAPORATION. 1972

Source: Periodic lake surface temperatures surveys. Evaporation computed from Richards and Webb, (1971,p.7), equation using Yellowknife Airport climatological data input. Includes two inches to cover evaporation during May and September

lake	Area (sq. mi. Approximate)	Surface Temperatures (° F)			Evaporation (inches)			Estimated Total Evaporation
		June	July	August	June	July	August	
Pocket	0.03	65	67	65	6.0	5.9	4.2	18.1
Vee	0.20	62	67	67	5.4	5.5	4.6	17.5
Stock	0.15	60	65	65	4.7	5.1	4.2	16.0
Frame	0.20	60	65	65	4.7	5.1	4.2	16.0
Kam	0.50	59	65	65	4.3	5.1	4.2	15.6
Long	0.25	58	65	65	4.0	5.1	4.2	15.3
Grace	0.20	56	63	65	3.2	4.3	4.2	13.7
Walsh	2.50	54	63	65	2.7	4.3	4.2	13.2
Prosperous	18.00	48	60	65	1.3	3.5	4.2	11.0

summer months for the medium-sized lakes in the sample are generally about five degrees higher than mean air temperatures for the same period. This trend is consistent with that reported by Webb (1973). The table shows that the smaller or shallower lakes such as Pocket, Vee, Stock and Frame warmed up relatively quickly in June attaining a mean temperature of 60°F or more in this month. The larger or deeper lakes such as Prosperous and Walsh lagged the smaller lakes by about 10-15° in June. The degree of lag between these two groups lessened considerably during July and almost vanished in August. It should be remembered that in all the summer months temperatures were above normal, including May, and were particularly so in June and August. It is therefore probable that the lake surface temperature would be a few degrees lower in the statistical 'average' year. This observation would also result in reduced evaporation in the average year by comparison with the pattern for the summer of 1972. The exact magnitude of the reduction is uncertain. It would however bring the seasonal totals of evaporation to levels closer to those found for the northern lakes in the Northern Ontario project (Richards and Webb, 1971). It is felt that even after a reduction for average conditions the temperature results from the present study would remain slightly greater. Part of the reason for this conclusion is that the lakes sampled in the present study were mostly much smaller than the Ontario sample mentioned in the literature review. This pattern of relatively small lakes is typical of almost the whole of Baker Creek basin. It may be remembered that the results for the northern Northern Ontario lakes corresponded well with those suggested by Ferguson, O'Neill and Cork (1970). This latter study claimed to present free water evaporation estimates for small lakes in which heat storage effects are small by comparison with direct, immediate heat exchange with the atmosphere. The Ontario lakes studied appear, in relation to the present study sample, to have been relatively large in size reducing the likelihood that heat storage effects were negligible. It is probable therefore that smaller lakes in Northern Ontario have greater evaporation than is suggested by the study and that true free water evaporation estimates for the region should be greater than is indicated by Ferguson, O'Neill and Cork (1970). Since the Northern Ontario region was mapped by the latter group to lie between the same isolines (12-16 inches) (Figure 3-1) as the present study area it seems reasonable to suspect that estimates for it will also be low. The results presented in Table 5-9 bear this suspicion out. Further support may be obtained from the Class 'A' pan results (Table 5-5).

The relatively rapid warming of the lakes early in the evaporation season appears to give support to the belief that solar radiation is an important factor in the heat balance of the lakes in the study area just prior to break-up. This would also support the concomitant contention that white ice or bubbly ice development on the lakes is not important and that the extinction coefficients of the ice covers is relatively low permitting the penetration of a large proportion of the solar radiation. This hypothesis needs further testing however since the warm conditions during May 1972 may have been a significant factor in the rapid warming.

The rapid warming pattern also contributes to lessen the likelihood of significant condensation influences on the larger lakes at least for this particular year. Condensation may be expected to be most favoured during the period late April to early June in the study area. At this time the large heat storage capacity of the larger lakes might be expected to contribute to slow the warming of the lakes and contribute to relatively low surface temperatures which approach the reigning dew point temperatures. The actual likelihood of this correspondence occurring may be guaged for June from Table 5-10 where mean dew point and air temperatures for four times of the day during the month over the past six summers are presented. The highest dew point temperature, 41°F, is well below the estimated June 1972 mean for the surface temperature of Prosperous Lake (48°F) which is the largest lake in the region. Even allowing for the fact that May and June 1972 were relatively warm it is unlikely that average temperatures would result in conditions favouring condensation. Baker Creek basin is dominated by lakes very much smaller than Prosperous Lake in size and it may be safely assumed that condensation onto these lakes in June at least is negligible.

TABLE 5-10

Mean dew point and air temperatures, for four times of the day. Yellowknife Airport. June, 1967-1972. (°F)

	M.D.T.			
June D.P. (Air temp.)	0500	1100	1700	2300
1967	36	36	35	37
	(44)	(52)	(57)	(48)
68	39	41	39	41
	(47)	(56)	(60)	(51)
69	32	31	29	30
	(44)	(53)	(55)	(47)
70	40	40	40	40
	(51)	(59)	(62)	(55)
71	41	39	38	40
	(53)	(63)	(66)	(57)
72	41	41	40	41
	(51)	(59)	(62)	(54)

Source: Monthly Record. D.O.T. Meteorological Branch. (Now Atmospheric Environment Service).

The June comparison has been performed because of the availability of surface temperatures for this month. Since the lakes in Baker Creek basin tend to open earlier than the larger lakes it is possible that they could experience noteworthy condensation gains in late April and early May. However they appear to warm up so quickly that significant gains would be precluded. This whole aspect needs much further investigation however utilising ground surface air temperatures instead of those from the Stevenson Screen level.

The Water Survey lake level records for the station on Pocket Lake (No. 07SB008) provide a means of estimating evaporation without having to resort to problematic empirical formulae and lake surface temperatures. The method is not without its own problems however. Assumptions have to be made with regard to an allowance for inflow into the basin from the surrounding land. (Pocket Lake basin is 36 percent lake surface). Similarly allowance has to be made for seepage out of the bottom or sides of the lake. The precipitation records for Yellowknife Airport have to be accepted as representative of Pocket Lake basin. In interpreting the lake level records an important point is to establish the overflow level and if possible to make an allowance for overflow if it occurs in a month for which an evaporation estimate is desired. With reference to this point it was not possible to determine the overflow level exactly. It is known from Water Survey records that no flow occurred from 1967 - 1969. Overflow is most likely to have occurred during breakup in 1970, 1971 and possibly 1972. However, because of gaps in the records the lake level peaks for these periods are not known. No flow was observed by this writer during field reconnaissance at the end of May, 1972. The channel immediately below the lake outlet showed signs of recent moderate flow conditions. A backward projection was made of the summer gradient expressing the fall in lake level from when records were available. This indicated that during the period of breakup the lake level was about 97.55 feet above the datum at Great Slave Lake (See Figure 5-8). In view of the evidence indicating slight overflow in the outlet channel it appears that this level is a close approximation of the overflow level.

Missing records preclude an evaluation of the influence of overflow amounts on the monthly water balance of the lake. This lack is not serious however since the influence may only be expected to be significant during parts of April and May in certain high level years and evaporation in these periods is probably relatively low. Lake levels for the beginning of June in every year were below the estimated overflow level, and hence no allowance has been made for its influence.

The influence of inflow into Pocket Lake from the surrounding basin will be conditioned by the amount, frequency and intensity of precipitation, the proportion of bedrock-dominated areas with low water retention capacities; the proportion of vegetated depressions with high retention capacities; the mutual distribution of the latter two surface types; and the intensity of evaporation from the basin.

During the period June to September the average precipitation is about five inches. Large deviations from this average may be expected in any year which may usually be related to a particularly wet or dry fall period. The precipitation is usually moderate in frequency and light in intensity, most events yielding less than 0.25 inches. The detention capacity of almost all surfaces in the basin can absorb such amounts very easily. However, in the abnormally wet months two or three intensive and high yielding storms will be responsible for over 80 percent of that month's precipitation. Such may be the intensity of the storms that detention capacities will be exceeded and runoff into the lake may be anticipated. In most cases there will be adequate capacity in the vegetated moss-tree covered depressions to absorb the highest yielding storms. These surfaces will also probably be capable of absorbing some runoff from adjacent bedrock dominated surfaces. As a result significant inflow to the lake may be expected only with relatively high yielding storms of 0.50 inches or more and only from bedrock-dominated surfaces immediately adjacent to the lake where no intervening muskegs may intercept the runoff. The area of bedrock surfaces draining directly into Pocket Lake may be estimated at approximately 10 percent of the total basin. It will probably vary greatly depending on the level and intensity of the precipitation events. Therefore, even for an intensive precipitation event yielding say an inch, over the basin as a whole, runoff into the lake might only add 0.10 inches to the one inch rise in level caused by precipitation directly onto the lake. Sprinkler runoff experiments conducted by A. G. Landals in July 1968 over very small basins in Baker Creek basin produced evidence indicating support for the comments made above (Personal Communication, June 27th, 1973).

High rates of evaporation from the bedrock surfaces are possible as a result of the dry, warm and breezy conditions which generally persist between the isolated precipitation events. As a result the detention storage of these surfaces are probably close to maximum levels just prior to the events.

This discussion has indicated that inflow into the lake from the surrounding basin will be very low, becoming significant only during the high yielding intensive storms. The latter are relatively rare in the study area, but in the odd month they can be very prominent eg. August 1969, 1970, 1971; September 1968, 1972 . A greater allowance for the influence of inflow into the lake would have to be made in these months whereas in most months the influence could be assumed to be almost negligible. In the following estimation of Pocket Lake evaporation from lake level records, inflow during relatively normal precipitation months was arbitrarily assumed to be 5 percent of the monthly precipitation. During months in which one or more storms occurred precipitating in excess of 0.50 inches the upward adjustment adopted was 15 percent of monthly precipitation. If over two inches of monthly precipitation appeared to be associated with only one or two storms the adjustment was +20 percent and +30 percent if over three inches occurred. The latter three classes of adjustment were noted to affect only 25 percent of the months surveyed. Appendix B was used as the guide for adjustments. The adjustment procedure could be further refined but a brief error analysis indicates that this would have only a small effect on the final evaporation estimate.

Seepage losses from the bottom and sides of the lake are probably very low since the lake is bounded on almost all sides by granite bedrock. It is also likely that the bottom of the lake and the parts of the boundary where rock is absent are plugged by fine glacio-lacustrine sediments. Such sediments were noted as comprising the banks of the artificial channel just below the outlet of Pocket Lake. It is therefore suggested that seepage losses are virtually negligible. The magnitude of the losses that might occur is probably equalled by the likely undercatch of precipitation by the rain guage at Yellowknife Airport and no adjustment has been made for the latter possibility in the evaporation calculations. If it is proved to be necessary however the summer evaporation total would be altered by a relatively small amount (approximately 0.50 inches in the average year).

In the light of the above discussion summer evaporation may be estimated from the following simple equation.

Summer Evaporation = Precipitation on Pocket Lake + Net change in level of lake + Inflow.

The equation was applied to data derived in part from the changes in lake level from 1967 to 1972 presented in Figure 5-8. The results of applying this equation to data for Pocket Lake over this period is presented in Table 5-11.

The table indicates large variations in evaporation from year to year particularly in August and September. Since these are the months when abnormally high rainfall has tended to occur during the past six years it is possible that the inflow adjustment has been too low for heavy precipitation months. The low August values in 1968 and 1969 can be explained in strong part by the relatively low mean monthly temperatures in those months. Also a disproportionately high number of days in these months were rainy with accompanying conditions which did not favour a high rate of evaporation. The year to year variability in the September totals can also be explained by large temperature and precipitation differences from normal, but the effect of snow precipitation and ice conditions in some years is an added complicating factor which could not be determined exactly.

It is believed that the total evaporation for June, July and August in each year represents a good estimate of evaporation for the main summer period. The six-year mean evaporation for this period is 11.03 inches. Total lake evaporation for any one year could be obtained by adding two to five inches depending on the particular early and late season conditions. In the case of 1972 with slightly above average temperatures in May and a very cold second half of September an increase of about 2.50 inches seems justified.

FIGURE 5-8: CHANGES IN LEVEL OF POCKET LAKE, 1967-1972
Source: Surface Water Data, Water Survey of Canada, Environment Canada.
N.B. Levels are in feet and refer to a datum at Giant Yellowknife Mine at the edge of Great Slave Lake.

Dashed lines represent gaps in the available record.

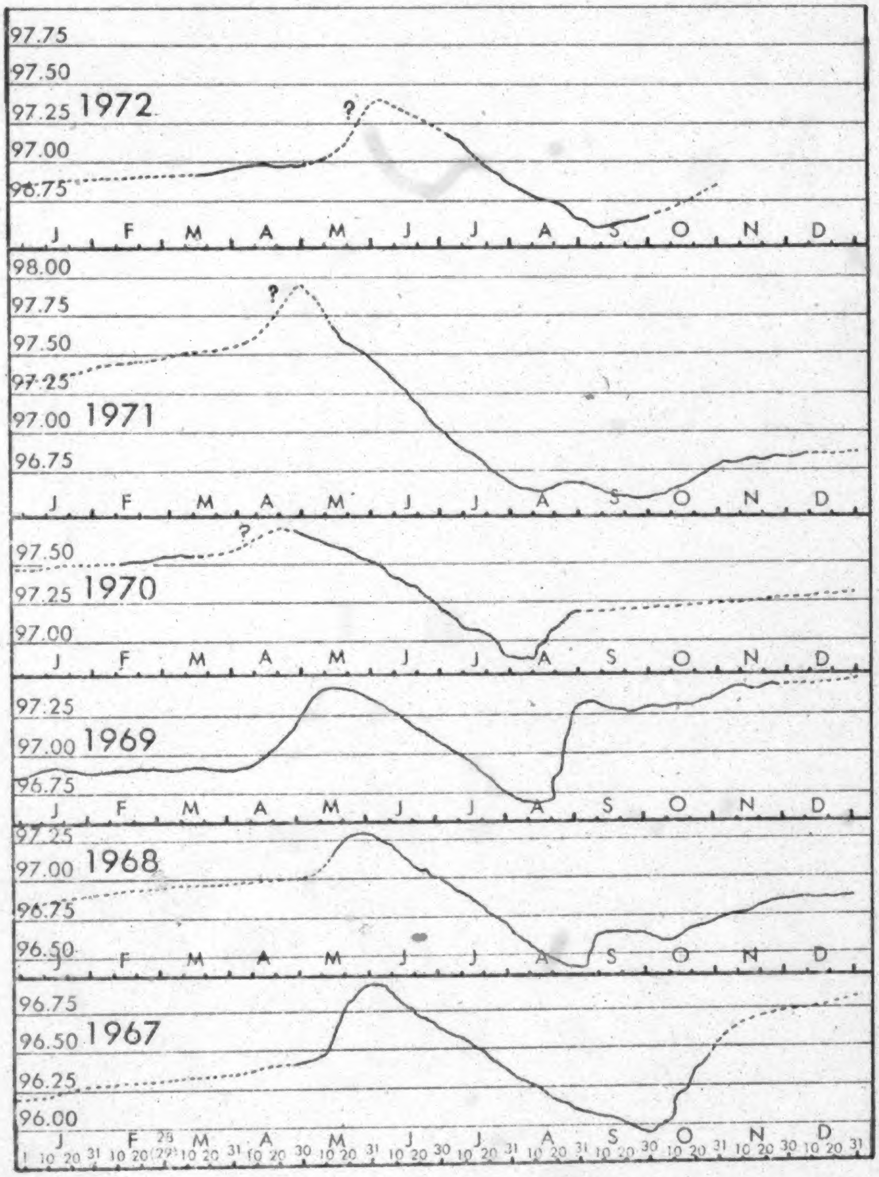


TABLE 5-11

MONTHLY EVAPORATION FROM POCKET LAKE ACCORDING TO LAKE LEVEL RECORDS.

(inches)

	June	July	August	Sept.	3 month total (J,J,A)
1972	3.97	5.88	3.86	ice problems	13.71
71	5.92	4.68	2.55	2.55	13.15
70	4.29	5.44	1.17	1.17	10.90
69	4.04	4.65	0.58	0.58	9.27
68	3.93	4.21	0.63	0.63	8.77
67	4.09	4.14	2.16	2.16	10.39
Mean	4.37	4.83	1.83	1.42	11.03

Source: Calculations by author; See text for method and assumptions employed.

This would result in a seasonal total evaporation for Pocket Lake of about 16.2 inches. Average year conditions would probably result in this total being quite close to the 14 inches published estimate by Ferguson, O'Neill and Cork (1970).

When compared with the evaporation totals derived from the lake surface temperature surveys and the semi-empirical equation the above evaporation estimates for Pocket Lake in 1972 are seen to be significantly lower. The June to August total by the former method was 16.1 inches compared with 13.71 inches determined using the lake level records. The latter results are believed to be slightly more accurate because of the greater scope for significant errors associated with the higher value. A prime factor in this respect is the probable unrepresentativeness of Yellowknife Airport wind speed and vapour pressure data for conditions over Pocket Lake. The steep, high rock walls bounding this small lake can be expected to shelter the surface from the wind. The airport site is on the other hand very exposed. The vapour pressure of the air over the lake is probably slightly greater on average than the air over the dry beach ridge airport site. There is also the possibility that the surface water temperature means were overestimated, again partly as a result of insufficient allowance for the depth factor. For its small size Pocket Lake appears to be quite deep although no precise observations are available. Allowance for these points in applying the formulaic method would result in lower levels of evaporation for Pocket Lake which would be closer to that determined from the lake level records.

The extrapolated mean monthly evaporation estimates provided by Ferguson, O'Neill and Cork (1970), Bruce and Weisman (1967), the Class 'A' pan 7 year mean values (Table 5-5), and the 6 year mean monthly evaporation estimated for Pocket Lake from lake level records are presented in Table 5-12. It should be noted that Pocket Lake is almost, but not quite because of its apparent relative depth, representative of small lakes with negligible heat storage. It is this category of lake only that is covered by the two published sets of estimates. Although the latter two sources indicate similar totals of evaporation the extrapolated monthly means possess different regimes. This reflects the attempt by Ferguson et.al (1970) to incorporate early season ice cover effects. The Class 'A' pan means are significantly higher, particularly in June and July, by comparison with the other estimates. It is indicated later that the pan data may reflect actual local advection effects but that the magnitude of the effect may not be quite so great as the pan data suggest. The published sets of estimates yield erratic comparisons with Pocket Lake data and it seems that neither can be relied upon as being truly indicative of small lake evaporation patterns in the study area. Some general comments may be offered however in the hope of fixing upon a satisfactory compromise position.

The evidence for relatively rapid warming of the lakes in the study region because of the influence of solar radiation

TABLE 5-12

ESTIMATES OF MONTHLY MEAN SMALL LAKE EVAPORATION FOR THE STUDY AREA

(inches)

	May	June	Jul	Aug	Sept
Pocket Lake	--	4.37	4.83	1.83	1.42
Class A Pan		6.21	6.36	4.32	1.80
Ferguson *et al* 1970	0.4	3.40	5.00	4.00	1.40
Bruce and Weisman 1967	2.8	3.50	3.40	2.30	1.40
Compromise position	1.40	4.50	4.90	3.20	1.40

Sources: See text discussion.

indicates that Ferguson, O'Neill and Cork (1970) were overcautious in allowing for ice cover effects. The estimate for May by Bruce and Weisman seems too high however since 2.8 inches of evaporation would, from the formula, correspond with mean lake surface temperatures for the whole of May of approximately 55°F although use of an average temperature in a month of rapid temperature change is questionable. This value seems too high by 5 to 10°F and allowance for this would lower Bruce and Weisman estimate. It is suggested that 1.40 inches would be a good compromise for May.

The Pocket Lake mean for June appears high by comparison with the published estimates but this situation can be defended on advection grounds. Much advective energy appears to be generated in the summer in the study area as a result of the low rate of utilization of available radiant energy by evaporation or evapotranspiration over the bedrock and some vegetative surfaces surrounding the basin. This point will be explained in a later section of this chapter.

The Pocket Lake July mean is probably the most accurate of the monthly estimates because it was derived from the section of the lake level records where possible overflow influences were definitely absent. Also July precipitation variations are much more conservative than is the case for August and September.

The Pocket Lake mean for August is very low by comparison with other estimates and it is possible that inflow influences during August 1969 and August 1970 when precipitation was abnormally high, have been underestimated. Ferguson *et al*'s estimate appears rather high on the grounds that available net radiation estimates for this month (Hay 1970) correspond to evaporating powers of just over three inches. It is therefore suggested that for August a more satisfactory compromise mean for small lakes in the study area would be about 3.20 inches.

During September the published sources and the Pocket Lake mean value agree best. It is therefore suggested that 1.40 inches is a good approximation for this month. This is about 0.50 inches above that likely according to Hay's net radiation estimation for this month. However the extra evaporation is justified on the basis of heat previously stored in the lake being released during this period, partly by evaporation.

The result of the compromise analysis has been added at the foot of Table 5-12. The annual total is 15.40 inches. Since Pocket Lake is quite deep and sheltered by comparison with most other small lakes in the basin it is suggested that true small lake evaporation in the basin might range between 16 and 18 inches although later evidence indicates that 18 inches is a plausible basin mean.

POOL EVAPORATION:

Data plots for the change in level of the pools are presented in Figure 5-9. Plates 5-2a and 5-2b indicate the changes in P1 over two months. It should be noted prior to interpretation that the measurements were not made at regular times through the day and that a few days were missed due to

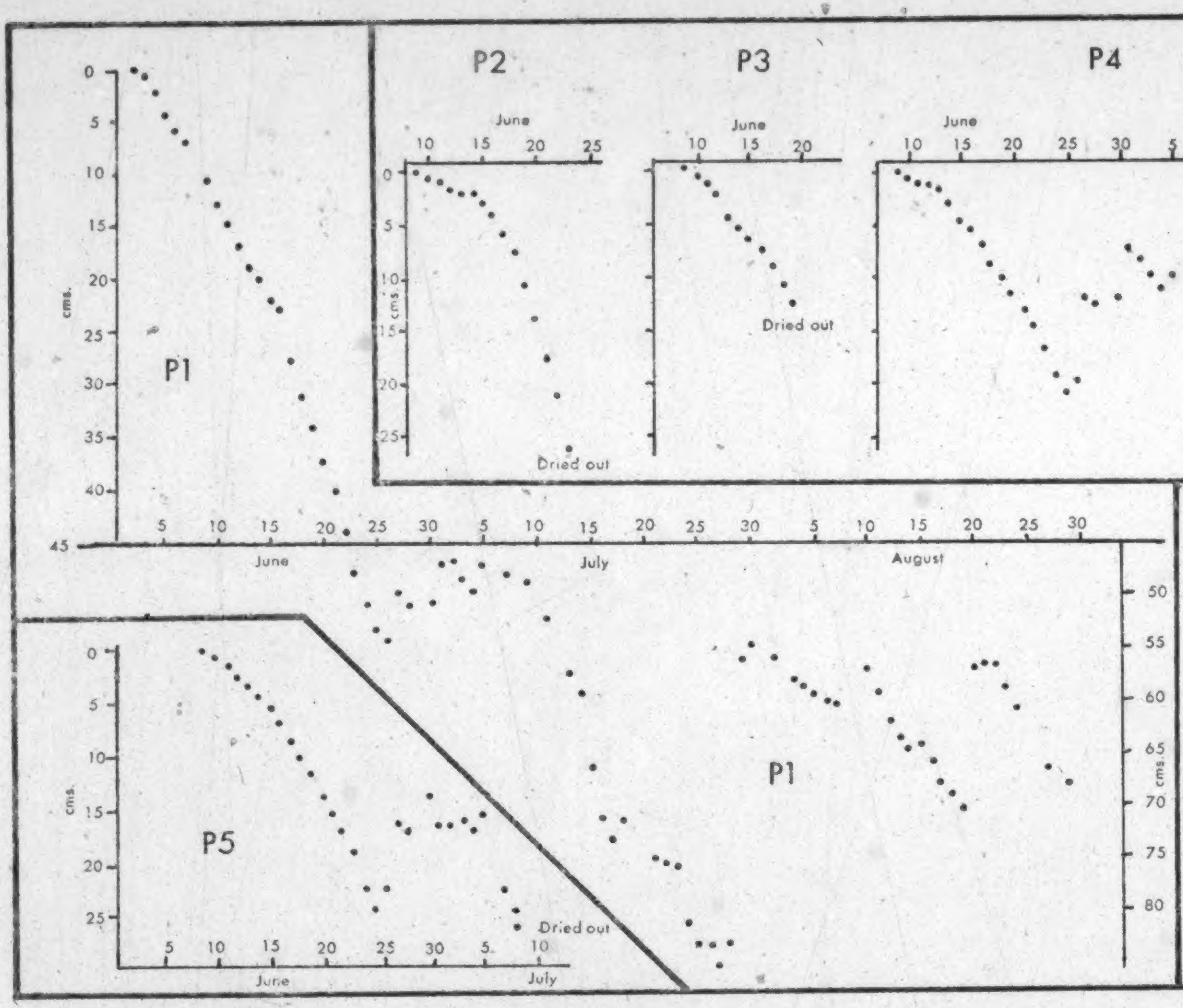


FIGURE 5-9.
CHANGES IN LEVEL
OF SMALL POOLS
STUDY AREA,
SUMMER 1972.

Source:

Field Observations.

Plate 5-2: View of part of Pool 1. 2a (June 3rd 1972)
2b. (August 3rd 1972).
See also Plate 4-12.
Photographs taken facing in a Southwesterly
direction.

the impassability of the road. This latter restriction applied usually after a significant storm so that the true magnitude of the storm's effect on level change may have been under-represented slightly. Therefore it is not valid to accept a line linking two plotted points as being indicative of the exact pattern of water level variation for the period. Despite these qualifications it is believed that the broad natural trends are faithfully reproduced for their present purpose.

Certain assumptions must be made before the plots can be interpreted with an evaporation focus. Seepage through the bottom of the pools is assumed to be very low or negligible. This seems to be a particularly safe assumption for P1 since there is a good possibility that the permafrost conditions present for the surrounding area extend under this large pool. During their existence the pools at the sedge meadow site are also underlain by permafrost at a shallow depth indicating negligible bottom seepage at these pools also. It is true that the sedge meadow is a drainage site raising the possibility of horizontal seepage. However the main drainage course through the meadow is about 100 metres distant from the sites and this fact tends to minimise the possibility of horizontal seepage.

The data plots provide some interesting information on moisture gradient and local advection influences in the study area.

In terms of moisture gradients it was interesting to compare the changes in level associated with a series of precipitation events. During the late June wet spell approximately 4.5 cms. of rain fell. The level of P1 rose on average by just over this amount. At P4 and P5 the levels rose by double the precipitation amount. The multi-peaking pattern associated with this series of rain events was accentuated at these two sites because of their small size by comparison with P1. Overall however it may be observed that levels rose in close sympathy with precipitation inputs during this period. It will be noted later that soil moisture tended to be well above field capacity during this period.

A different pattern occurred in association with the rain event on July 28/29 when approximately 2 cms. of precipitation fell in a period of about twelve hours. The meadow pool sites had dried out by this time and no effect of the heavy precipitation was observed. Soil moisture levels were well below field capacity at this time and were able to absorb the precipitation readily. At P1 the 2 cms. of precipitation produced a 30 cm. rise in level of the lake indicating considerable inflow from an extensive pool catchment area. Just prior to this storm P1 reached its lowest level of the season, approximately eight feet below initial levels. Moisture gradients were therefore well developed and it appears that surplus moisture travelled along them rapidly to the pool.

During a rainy period from about August 7-9th 0.50 cms. of precipitation resulted in a 3 cm. rise in level. It is reasonable to expect less well-developed gradients at this time since the level was not as low as just prior to the late July storm. Lower levels occurred just prior to the August 19th rain event however when one cm. of precipitation fell and P1 rose by about 12 cms.

These observations point to the existence of well developed moisture gradients under the bog surfaces surrounding the lakes in the latter part of the summer. It is possible that this situation was favoured by the silty, sandy mineral soil conditions which were observed at the bottom of the pool as it dried out. The exact depth relations of the organic and mineral soil layers under the surrounding bog surface is not known. It is probably in order however to caution against extension of the results from P1 to sites where the organic layer is dominant and where less efficient moisture gradients may be expected to develop.

It was possible to deduce tentative evaporation rates from the pools during periods when precipitation was absent or negligible. The results of this procedure are summarised in Table 5-13. The quality of these results depends on the negligible seepage assumptions. If these assumptions are valid the rates shown in the table are significantly higher than those which may be estimated from available net radiation data (Hay 1970). These latter data indicate that approximate evaporation rates for the pools would be: June 0.50 cms. per day, (0.20 inches per day); July 0.40 cms. per day, (0.15 inches per day); August 0.30 cms. per day, (0.12 inches per day). These rates imply that substantial local advection

TABLE 5-13. TENTATIVE EVAPORATION RATES FROM POOLS FOR SELECTED PERIODS.

Summer 1972.

	EVAPORATION RATE			EVAPORATION RATE	
P1	(cms./day)	(inches/day)	P4	(cms./day)	(inches/day)
June 2-26	2.5	1.00	June 9-25 overall	1.3	0.50
July 9-27	1.8	0.70	June 9-17	0.9	0.40
July 30-7 August	0.6	0.20	June 17-25	1.8	0.70
August 10-18	1.4	0.60	P5		
August 21-28	1.6	0.65	June 9-25 overall	1.5	0.60
P2			June 9-17	1.0	0.40
June 9-23 overall	2.0	0.80	June 17-25	2.0	0.80
June 9-16	0.7	0.30	P3		
June 16-23	3.1	1.20	June 9-19	1.3	0.50

Source: Estimations by author from data presented in Figure 5-9.

influences are present for the pools, particularly Pl.

It may be noted that, during June, rates of evaporation for the sedge meadow pools are about half of those found at Pl. This may be a logical function of the greater size and exposure of Pl. However it is possible that seepage is an important factor in causing the large falls in level from which the evaporation rates were inferred. The mineral soil conditions present at Pl increase the likelihood of this being the case, especially if permafrost conditions do not extend below the pool. Even if seepage is important however the discrepancy between rates corrected for seepage loss and the net radiation equivalent rates would still be significant. Local advection may therefore well be an important factor for consideration.

Local advection may be considered as the exchange of energy and moisture due to horizontal heterogeneity of the surface. Philip (1959, p 535) has noted that the neglect of advection has tended to vitiate many attempts to treat the problem of natural evaporation. He believes there is a logical flaw in treating the humidity and temperature of the air solely as the cause of evaporation at the underlying surface. To him they are equally the consequences of the upwind evaporation and energy conditions. Rider and Philip (1960) indicate that the most important factor operating in advective evaporative effects is the change in the partition of available energy between the sensible heat and latent heat term. In other words, wherever advection arises from changes in availability of water for evaporation it is attended by a sudden (spatial) change in the Bowen ratio.

It is believed that in the present study area, because of the intricate mosaic of surface relations, local advection effects could be a significant aspect of the water balance. This belief is accentuated by the very low levels of water availability indicated for bedrock dominated surfaces for much of the summer period. Also, results from a later section indicate that water availability through many of the vegetated surfaces in the region is low despite moist underfoot conditions.

Rider, Philip, and Bradley (1963) have reported a critical micrometeorological study of advection influences over a discontinuity between a dry paved concrete runway and a well-watered grass plot. They derived a quantity, R/Lp.w., which was a generous upper limit to the evaporation rate from a large area free from advective effects exhibiting the general properties of the grass plot under the prevailing radiation balance. This quantity was in effect an estimate of the rate which would have occurred under the same conditions from an area of infinite extent. They derived a set of ratios which gave an index of the magnitude of the evaporative advective effects, integrated for certain distances from the discontinuity, in relation to R/lp.w.: This set of ratios was

$E_{0\text{-}100 \text{ cms.}} : E_{0\text{-}400 \text{ cms.}} : E\ 0\text{-}1600 \text{ cms.} : R.L.p.w. = 3.7 : 2.5$
$: 1.7 : 1$

(Rider, Philip and Bradley 1963, p 529).

The authors felt that this was a conservative index. Therefore it is likely that the boundary zone experiencing advective effects could be greater than sixteen meters. It is possible that similar boundary zone influences occur in the present study area. Since the lake distribution is one of very many small lakes rather than a few large ones the potential for this influence being significant is increased.

These comments must be taken more as hypotheses to be tested by future research rather than established patterns. The argument does however appear to explain in large part the high rates of evaporation apparent at P1. Also the moisture gradient conclusions suggest that availability of water under the bog vegetation surrounding the pools is particularly limited.

NET RADIATION

Daily radiation totals obtained from the net radiometer at the Muskeg site (YT2) are presented in Table 5-14. The accuracy of these values are in doubt to the extent that no correlations with other parameters have been attempted. In the analysis use was made of the more reliable monthly means of daily net radiation. These are presented in Table 5-15 with other available estimates for these months. It may be noted that the results obtained in the present study compare quite well with published estimates. Since the latter are annual averages it would be helpful to have some estimate of the representativeness of the results for the summer of 1972. It has already been noted that this period was warmer than

TABLE 5-14.

DAILY TOTALS OF NET RADIATION AT YT 2.

June-August 1972.

(Cals. per sq.cm.)

Day of month	June	July	August
1		156	94
2		324	228
3		263	186
4		304	226
5		237	125
6	installed	309	205
7	380	260	173
8	385	355	188
9	291	222	58
10	390	250	197
11	326	283	152
12	195	287	173
13		267	153
14		300	104
15	308	302	168
16	330	230	160
17	318	293	159
18	283	269	152
19	306	205	32
20	351	240	154
21	310	214	168
22	298	244	137
23	314	236	95
24	209	213	124
25	76	234	100
26	233		78
27	268		75
28	96		
29			192
30		195	163
31		263	229

Source: Field Measurements.

TABLE 5-15

Selected estimates of monthly
mean net radiation (cals. sq. cm./day)
for the study area. June - August.

Wendland (1967)	June	July	August
1960	M	M	140
1961	232	235	135
1962	355	267	192
1963	306	280	216
Mean (1960-1963)	298	261	171
Wendland (1965) reported in Hay (1970) with 35 cals. sq. cm./day added for regional representativeness.	335	295	215
without +35	300	260	180
Hay (1970	296	249	153
Budyko (1963)	277	226	161
Present study (YT2)	282	258	148

M = Missing Data

Sources: As Referenced.

TABLE 5-16

Sunshine data. Yellowknife Airport 1970-72

		Duration in hours	Percent possible duration
1970			
	June	353	60
	July	359	62
	August	250	51
1971			
	June	427	73
	July	395	68
	August	267	54
1972			
	June	376	64
	July	415	72
	August	291	59

Source: Monthly Record.

normal, however temperature is not a good sole indicator of net radiation patterns. Sunshine records are available for Yellowknife Airport only since 1970 (Table 5-16). The data indicate that the summer of 1972 was relatively sunny, particularly in July and August. It is possible therefore that in the average year the values could be slightly lower than in 1972 for July and August and higher for June. Such adjustments would result in values which are most consistent with estimates by Hay (1970) rather than those by Budyko and Wendland.

It is important to be aware of the merits and demerits of the published estimates of net radiation. Comments here are confined to an evaluation of the estimates by Hay and Wendland. It is anticipated that in general comments on the research by Hay also apply to that of Budyko although the latter study was on a global or hemisphere scale compared to the regional scale of Hay (Hay, 1970, p 22).

Hay (1970) derived his estimates by means of a computational model, using climatological data to compute the radiative fluxes for a particular location. His use of monthly mean values assumed no excessive biasing in the results. He was able to prove this assumption to be valid for Toronto Scarborough where records were best suited for making the check. It is possible that in a contrasting climatic situation such as for the present study area that the assumption might not be valid. Because of constraints on the climatological data acquisition Hay calculated his values from data for the period 1957-1964 inclusive. If this period was not representative of average conditions in the present study area then Hay's estimates for this location would not be suitable as means. There is also the question of the representativeness of Yellowknife Airport data for the surrounding terrain. For certain model inputs such as sunshine and cloudiness factors this question is probably not serious; however spatial air and surface temperature variations may well be significant and affect the model estimates.

Hay (1970, p 160-62) has attempted his own evaluation of the difference between his own and Wendland's estimates as noted earlier. Hay's argument in favour of his own estimates is accepted although reservations remain which may only be removed by detailed studies involving simultaneous measurement of net radiation at the Airport site and at sites in the basin. If such studies were coupled with a re-evaluation of Wendland's data a wealth of information on net radiation patterns in the study area could be obtained.

In his comparison Hay added to Wendland's first estimates, (Wendland 1965), a correction of 35 cal. $cm.^{-2}$ day^{-1} suggested by Wendland for regional representativeness because of differences in albedo of the surface over which the radiometer was placed and the albedo over the surrounding Boreal Forest. This procedure is probably of questionable validity since Hay's own estimates are based on Yellowknife Airport data, the regional representativeness of which is not established. However, Hay noted that even when this correction is omitted the Wendland estimates still indicated higher values of net radiation.*

The results obtained in the present study have not been corrected for the beam radiation reflection problem which could entail an upward adjustment up to 15 percent in the mean values. The practice of not correcting for this problem follows that of Wendland (1967, p 3) who felt that because of the significant range of accuracy of the instrument the additional bias contributed by the reflection problem was not worthy of consideration. If incorporated results for both the present study and Wendland's would be increased.

The measurements made by Wendland (1967) were "supposedly made over a somewhat sandy surface with a few tussocks of grass." (Wendland, Personal Communication, April 21, 1973). His values which exclude the regional representative correction, (and an upward adjustment for the reflection problem), are representative of a surface with a relatively high albedo. Geiger (1965, p 15) has indicated an albedo range of 15 to 40 percent for sandy soil. For the dry conditions which predominate during the summer months at the Yellowknife Airport site it is likely that the albedo will average about 25-30 percent, which is probably the highest albedo likely to be

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* There is some discrepancy between the estimates of Wendland (1967, p 9) and Hay's version of these estimates. Hay appears to have rounded off the total obtained by adding the correction factor. In the case of the August value an arithmetic error appears to have been made. Therefore Wendland's own uncorrected estimates should be used.

experienced in the region during this period. It is therefore logical to expect relatively low net radiation measurements over this site. Such results would have been realised if Wendland had not omitted "many five to six hour segments.... due to weather conditions " (1965, p 6), when inclement weather conditions persisted. It is probable that actual net radiation would have been low enough to absorb the regional representative factor, and any allowance for the reflection problem, to such an extent that resulting values would more closely approximate those derived by Hay.

The above reasoning suggests that the Hay estimates are good indications of spatially integrated net radiation patterns in the region. Later reasoning, on albedo grounds suggests a similar conclusion from results obtained in the present study. It may also be noted that Hay found his results to be compatible with those of Ahrnsbrak (1968). Therefore in the absence of strong evidence to the contrary it may be assumed that Hay's estimates are valid regional approximations of net radiation.

In the present study the net radiation was measured at the muskeg site. The representativeness of the results gained there in relation to the other surfaces of interest may be gauged from a consideration of the subarctic albedo patterns determined mainly by Davies (1962).

The study site has been burned over, probably about ten years ago. A few charred deadfall may still be observed. The dead trees have been stripped of their bark and reflect strongly, this being particularly noticeable under direct sunlight. Also many dead trees or branches litter the ground providing good reflecting surfaces for radiation perpendicular upon them. The green shrub layer has not regenerated sufficiently to reduce the exposure of the deadfall (See Plates 4-1, 4-2, 4-3). As a result of the burn characteristics the albedo of this site is relatively high by comparison with most muskeg surfaces. Another possible contributing factor to the high albedo is the comparatively well-drained conditions.

According to Davies (1962, p 37-8, 83) the mean representative value for the albedo of burns is 12.4 percent. Use of this highly specific value is not without risk since a wide range occurs (See Table 3-8). It is accepted however since the wide range affects the other sites and averages for these have been considered also for ease of analysis in the absence of detailed information. Bog, muskeg and water surfaces range below this on average while the albedo for the bedrock dominated surfaces (the 'tundra' cover type) ranges to about the same magnitude above the burn value. This would indicate that the burn site may provide a good estimate of regionally integrated net radiation. Since the adjustments to the 1972 measured values to represent average conditions resulted in estimates most consistent with those published by Hay, this would indicate that Hay's results are good regional integrations of net radiation.

Other sites of interest in the present study were water surfaces, spruce bog and sedge meadow. On an average basis water tends to have the lowest albedo of any surface during the summer period. It is certain therefore that net radiation at water surfaces was greater than the measurements from the muskeg site. Rouse (1973) has noted that net radiation for a lake was about 10 percent greater than a lowland swamp and about 25 percent greater than for an upland ridge. The lowland swamp type was made up of various bog and muskeg terrain whilst the upland ridge type comprised lichen woodland/open tundra lichen which has a relatively high albedo, averaging 22.1 percent. On the basis of these differences it may be suggested that net radiation for water surfaces is about 10-20 percent greater than that measured for the burned over muskeg.

The main spruce bog investigation site was not typical of the surrounding woodland. This does not appear to be critically important since according to Davies (1962, p 83) the representative value for bog and muskeg, and closed crown forest are quite similar, (9.7 percent and 9.2 percent albedo respectively). It may be anticipated that net radiation at this site averaged about 5 to 15 percent greater than that recorded at the burn.

Davies (1962) has provided some estimates of albedo derived by ground measurements which indicate that vegetation growth on a sedge meadow (and bogs in general) will shade surface water and contribute to raise the albedo. He indicated a mean albedo of 16.2 percent for the fringe of the sedge meadow (1962, p 78). It is possible that Davies may have overestimated the shading effect. Brown (1965, p 22) observed that net radiation values for sedge were higher than the darker coloured moss surface on clear days. This suggests the occurrence of wet patches on the surface unshaded by the vegetation. A similar situation is possible for the sedge meadow investigation in Baker Creek since many pools or wet depressions were present during much of the summer period. The values from YT2 may therefore be applicable to the sedge site with little or no adjustment.

The adjustments suggested above should be applied with caution since they are within the range of error anticipated for the measurements. The direction of the adjustments is probably valid but the magnitude is suspect.

It is regrettable that so much emphasis has had to be placed on the literature in order to derive conclusions for the net radiation aspects of the study. It is particularly regrettable since the original research design envisaged that field measurements of net radiation patterns over different surfaces would be the major thrust of the work in recognition of its fundamental importance in evapotranspiration terms. Severe instrument problems drastically curtailed these plans however causing a re-orientation of the thesis on much broader lines. Repair and recalibration of the instruments would permit another attempt at measuring the net radiation patterns in the basin at a later date. This would be most desirable as a check on the inferences drawn from the literature. Later in this chapter an attempt is made to use information from this section to derive crude monthly Bowen ratios according to the evapotranspiration losses from the sites.

Hay's estimates of net radiation for the basin are considered to be the best available. If the months of positive radiation balance are considered the potential evaporating power of the mean monthly net radiation in inches is: April 1.4; May 4.7; June 5.9; July 5.2; August 3.1; September 0.9, (Calculated from Hay 1970, p 160). These values must be reduced to allow for snowmelt and positive ground heat flux energy utilization. Snowmelt would constitute a moderately significant energy sink in the spring months for a short period only but ground heat flux is significant in other months, perhaps reducing the evaporating power by about 15 percent.

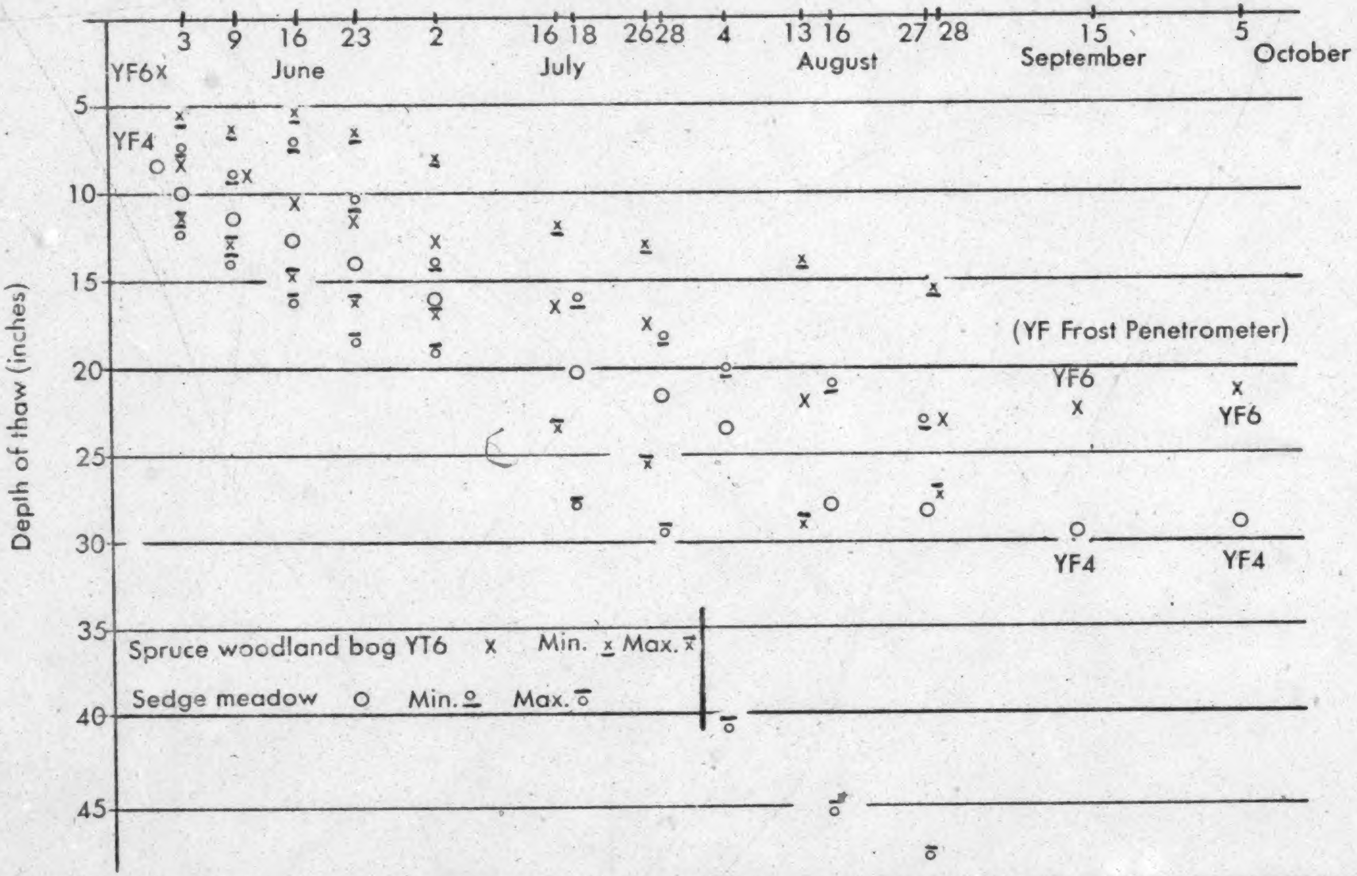
ACTIVE LAYER DEVELOPMENT

Some general information on ground heat exchange patterns may be derived from the results of monitoring the development of the active layer at sites underlain by permafrost. The results of the probe measurements at the sedge meadow and spruce bog sites are presented in Figure 5-10. In addition to the plot of the 25 point means for each measurement period the extreme deep and shallow observations have also been recorded.

The shallowest observations at both sites occurred in the dense woodland or ground birch microsites where shading of the surface is most pronounced. It is probable that the retarding influence of shading on active layer development is reinforced, and perhaps superseded by the insulating effect of the thick, loosely compacted moss cover which is also associated with the shading vegetation. (Brown, 1965, p 23). At the sedge

FIGURE 5-10. ACTIVE LAYER DEPTHS.YT4, YT6. SUMMER 1972.

Source: Field Measurements.



site the Bryum moss tends to be relatively thin and well compacted, resulting in less efficient insulation. The combined influence of better shading and more efficient moss insulation provides a likely explanation for the consistently shallower active layer development at the spruce bog site. It may be noted that this difference would have been accentuated if more emphasis had been given to the densely wooded microsite which is in effect most representative of conditions in the surrounding area.

The deepest observations were consistently recorded under the pool microsites indicating the significance of the abundant water as a heat conductor and heat storage medium. The sites intermediate between the influence of good shading and moss insulation, and highly saturated conditions are well represented by the plotted means. Results from frost penetrometers installed at the sites in connection with the ground temperature studies were found to be good agreement with the mean values. Readings from these sites during mid-September and early October have been included in Figure 5-10 in order to indicate the fall trend. Data for the period at the commencement of thawing are not available but it is probable that this occurred about mid-May.

The measurements of the depth of active layer development can be used to make a rough estimation of energy withdrawn from the available net radiation for this purpose. This estimation is aided by the high volumetric moisture contents present at the two sites under consideration. Monthly estimates of the energy consumed by the melting out of the active layer at each site are presented in Table 5-17. The volumetric soil moisture is an approximate mean value abstracted from data shown later (Figure 5-12). In the case of the spruce bog site the values are those for the main site (YT6M). The depth of active layer melt was obtained from Figure 5-10 by subtracting the depth at the beginning of the month from that at the end. Because of the high moisture contents it was assumed that the energy required to thaw the ground would be closely approximated by the energy required to thaw the water in the total moss. This can be obtained easily since the latent heat of melting of water is 80 cals. per gram. The heat necessary has been expressed as a percentage of the net radiation measured at the muskeg site for the months in question. It is assumed that these results approximate the net radiation at the two permafrost sites although it is probable that more radiation is received at the spruce bog, thus reducing the percentage of net radiation values slightly.

The results presented in Table 5-17 may be on the low side if allowance is made for the small amount of heat necessary to thaw the organic constituents of the moss. On the other hand the results may be too great if the statement of Nakano and Brown (1972) applies to the study sites. They have noted that: "It is well known that substantial amounts of water remain unfrozen at the freezing point of water in five grained soils" (Nakano and Brown, 1972,p20). They also mention the results of an experiment where more than 50 percent of the

TABLE 5-17

CRUDE ESTIMATION OF ENERGY CONSUMED IN THAWING OF THE ACTIVE LAYER AT YT4 AND YT6

(June, July, August, 1972)

	Soil Moisture (monthly mean Volumetric %)	Depth of Melting inches	Depth of Melting cms.	Energy used Cals. sq. cms.	Percent Net Radiation (YT2) (Approximate)
YT4					
June	85	8.0	20.5	1394	16.5
July	80	6.5	16.5	1056	13.4
August	77	6.0	15.0	924	18.0
YT6					
June	82	5.5	14.0	918	11.0
July	82	6.5	16.5	1082	13.5
August	78	4.0	10.0	624	13.5

Source: Author's calculations. See text for explanation.

'Energy used' calculated by multiplying together the soil moisture value (expressed as a percentage), the depth of melting (in cms.) and the value for the latent heat of melting of water (80 cals./gm.).

eg. YT4 June $\frac{85}{100} \times 20.5 \times 80 = 1394$ cals./sq.cm.

The net radiation values for YT2 were obtained from Table 5-5 (Present study YT2). The soil moisture values were obtained from Figure 5-12.

moisture contained in organic soil remains in the liquid phase in the temperature range down to -10°C (p 24). The implication of these statements is that the table results should be reduced quite significantly.

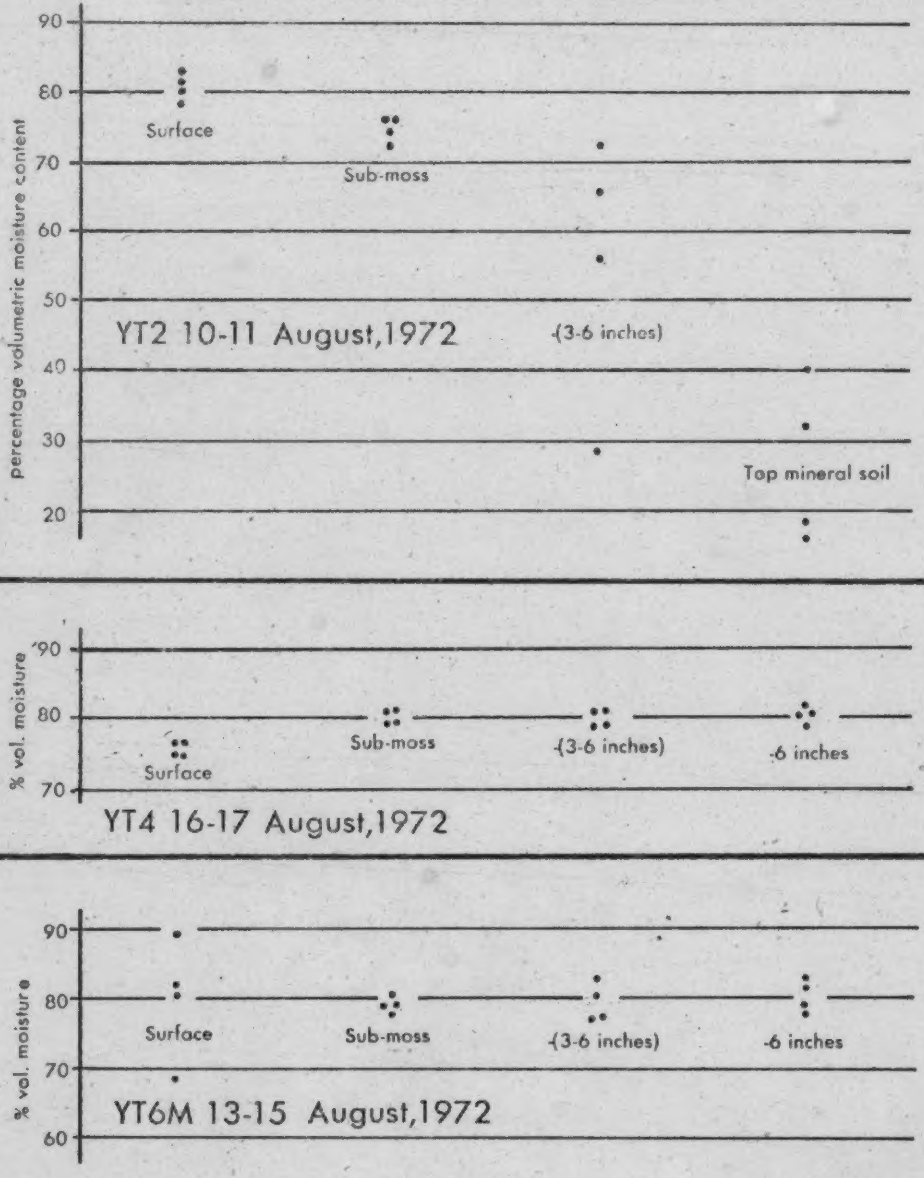
Once the moss has been thawed energy is utilised in warming the ground to temperatures above the freezing point. The importance of this component of ground heat flux is not known for the present study and it has been gauged from the literature survey. A reasonable average figure for total ground heat flux during the main summer period at the sites would be about 51 percent of available radiant energy. A lower value is perhaps more likely for the muskeg site where no permafrost but a significant mineral soil horizon exists.

SOIL MOISTURE

The results of the field capacity studies are presented in Figure 5-11. It may be noted that apart from some minor surface variations the field capacity values for the bog and sedge meadow are very similar. At the muskeg site the influence of the sand-silt mineral soil becomes more important with depth and results in lower field capacity values being realised by comparison with the other two sites.

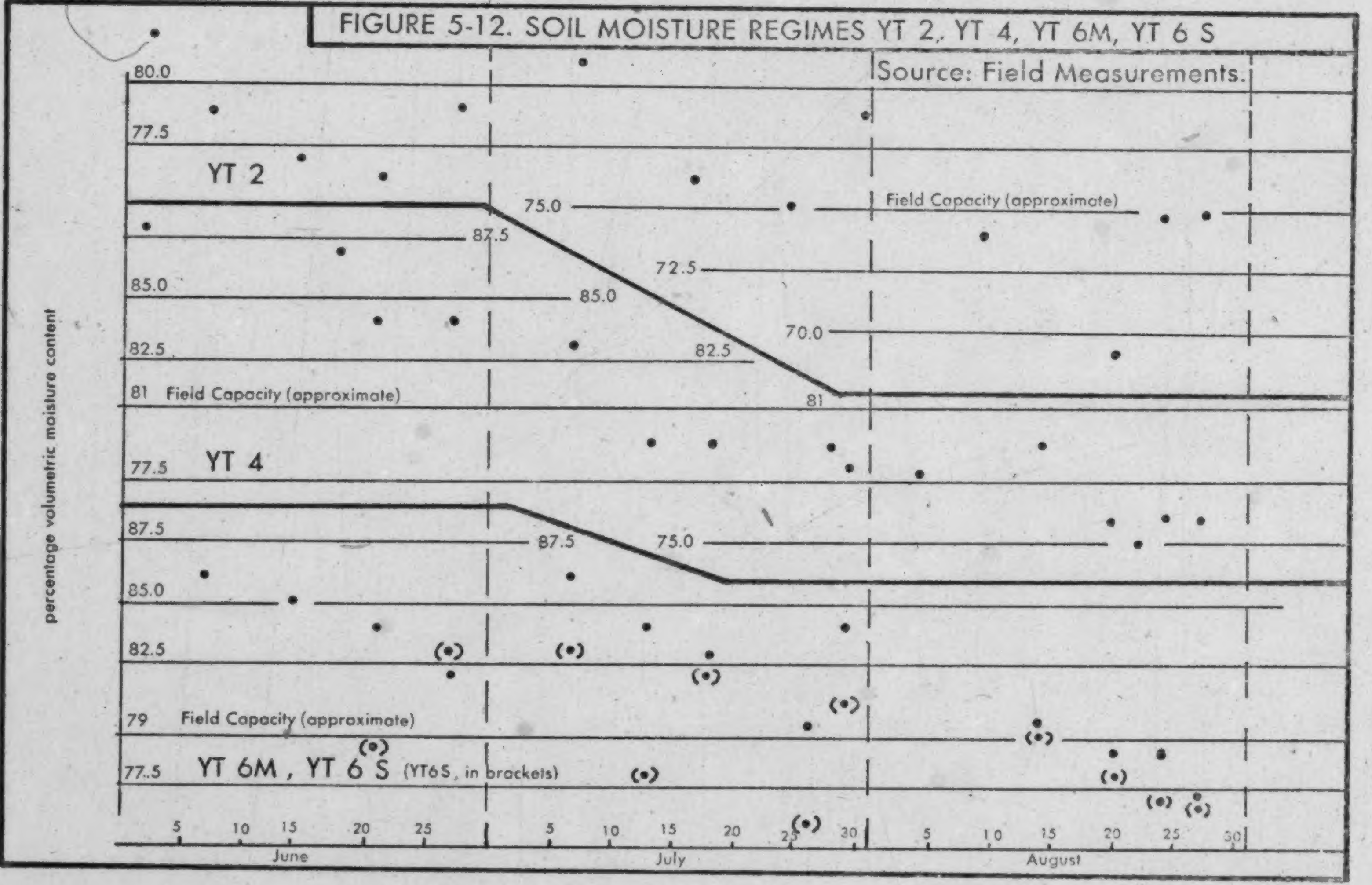
The summer soil moisture regime for each site is presented in Figure 5-12. The approximate field capacity values for the top layers of the soil at each site have been included. It should be stressed that it is not admissible to link up the points since precipitation events and drying periods occurred between measurements introducing additional fluctuations.

FIGURE 5-11. FIELD CAPACITIES OF THE SURFACE LAYERS.
YT2, YT4, YT 6 M.
N.B. Four moisture samples from each layer



Source: Field Measurements.

FIGURE 5-12. SOIL MOISTURE REGIMES YT 2, YT 4, YT 6M, YT 6 S


Source: Field Measurements.

it is believed to be unlikely that such fluctuations would differ markedly from the range indicated by the plots.

The highest soil moisture values were obtained at YT4 during June which is also the period when pools were most common over its surface. This site appears to dry out rapidly however by mid-July. A significant precipitation event of about 0.80 inches (2 cms.) in late July did not result in rejuvenation as did similar precipitation events in late June.

At the bog site, YT6M, the soil moisture remained slightly above field capacity for most of the summer although a definite downward trend, also found at the other sites can be detected. Some plots for YT6S have been included which indicate somewhat lower moisture values compared with the exposed level site. This difference may be due to the microrelief at YT6S which results in relatively dry hummock areas accounting for much of the effective evaporating surface, intermixed with very wet valleys.

At YT2 soil moisture values were slightly lower than at the other two sites, probably reflecting better drainage conditions, but for most of the summer period field capacity appears to have been exceeded.

The high frequency with which moisture contents exceed or approximate field capacity at the study sites may be noted. When such moisture conditions are met at mid-latitude sites under a normal vascular vegetation cover evapotranspiration occurs at the potential rate. This rate is normally approximated by the evaporating power of the net radiation.

The regimes indicate relatively little change in soil moisture in the course of the summer. A slight downward trend can be detected for each site but this in effect amounts to only a small fraction of an inch of water lost for every inch of soil over the three month period. Summer precipitation is generally well below that required to meet P.E. demands. The soil moisture necessary to make up the deficit could not be wholly supplied by the top few inches of the soil, on the basis of the degree of change monitored, but must be derived from deeper parts of the soil. As will be shown later an appreciation of this fact permits the derivation of adjustment factors for the evaporimeter results. These are necessary since the cans are relatively shallow and prevent the upward supply of moisture from deeper layers from taking place.

The samples taken from the pits dug to determine soil moisture/depth relationships indicated that there was little variation in soil moisture with depth at YT4 and YT6M. Generally the value for the top 5 inches (12.5 cms.) could be taken as a fair representation of the whole profile. Individual values deviated from this average by up to 10 percent in the case of YT6 and about 5 percent in the case of YT4. However the deviations showed no consistent pattern with depth. At YT2 the organic surface layer contained much higher soil moisture levels than the deeper mineral soil. Values for the latter ranged below the field capacity value determined for the mineral soil indicating relatively good drainage at this site at least after the seasonally frozen ground had thawed out. The presence of the permafrost table at the other two sites contributed to impeded drainage conditions.

It was not possible to obtain detailed information on the influence of thawing out of the active layer and the seasonally frozen zone on storage characteristics and runoff patterns. Little definite information is available in the literature on the influence of these factors in the present study area. Sommer and Spence (1968) suggested that the regime of the Yellowknife River and other rivers in permafrost areas could be explained by the release of moisture upon melting of the active layer in response to the summer temperature regime. Landals (1971) disputed this by maintaining that the nature of the regime is best explained by reference to the delaying effect of the lakes and other depression storage. MacIver (1971) analysed some of the 50 foot ground temperature profile data available for sites within the basin. For the permafrost sites, (YT4, YT6) he found that the active layer thawed to a generally constant depth such that a contribution later in the summer to annual streamflow discharge could not be expected from such areas. However for the muskeg site (YT2) MacIver found that once the cumulative degree-days over 32°F reaches approximately 3700°F there is an associated increase in the rate of depth of thaw. He noted that this degree-day total tended to occur in late August and suggested that the increased streamflow discharge in the Yellowknife area at this time of year appeared to be correlated with this marked increase in the depth of thaw of the muskeg. The proof of this relationship is confused by the tendency in recent years for heavy rainstorm events to occur in the months of August and September.

The present research indicates no noticeable effect of released moisture causing increased runoff. Summer base flows tend to be very low (Figure 5-7). Subsurface runoff is not favoured by low topographic gradients and the low water transmissability rates associated with peat. It could be checked however by means of piezometric transects across possible outlet zones. Experience with the evaporimeters, reported later, indicates that the upward flow of moisture from lower soil layers is a significant source of supply for evaporation from the bog and sedge meadow sites. On the other hand good drainage conditions were observed at the muskeg site once the seasonally frozen zone had thawed out thus counteracting upward flow potential, reducing moisture supply and reducing evapotranspiration despite the existence, as the soil moisture results show, of a very moist surface organic layer. It is conceivable that sites such as the muskeg (YT2) make up a good part of the baseflow in the latter half of the summer.

EVAPORIMETER STUDIES

A relatively direct estimate of evaporation/evapotranspiration for the muskeg, sedge meadow and bog sites was obtained by means of the batteries of evaporimeters. The detailed records obtained from the evaporimeters may be found in Appendix D. In view of the questionable reliability of the evaporimeter results over short measurement periods the data presented in Appendix D have been used to derive monthly estimates of unadjusted water loss at each site for June, July and August 1972. These are presented in Table 5-18.

The results shown in Table 5-18 indicate very low values of water loss by comparison with the values that might be expected from net radiation and soil moisture considerations. This observation casts doubt on the ability of the evaporimeters to adequately simulate natural conditions. It was felt that the main deficiency of the evaporimeters related to their inadequate representation of natural soil moisture conditions. It was found possible to guage the magnitude of this deficiency by utilising the soil moisture regime measurements and the changes in the weight of the evaporimeters over certain periods.

The evaporimeters contained about one to two inches of surface vegetation and about five inches (12.7 cms.) of moist soil. The soil moisture measurements were made from samples taken in the same five inch deep zone. By noting the volumetric soil moisture content at the beginning and end of a certain period it was possible to estimate the net change in soil moisture in inches through this zone. The change in weight of the evaporimeters in inches of water should correspond with that determined from the soil moisture measurements if the evaporimeters simulate natural soil moisture conditions exactly.

If the change in soil moisture is less than the change in

TABLE 5-18

Unadjusted monthly estimates of
water loss from evaporimeters at
YT2, YT4, YT6M, YT6S. June - Aug 1972
(cms.)

(precipitation in brackets)

		YT2	YT4	YT6S	YT6M
		3.90 (1.70)	4.40 (1.90)	3.70 (2.00)	7.40 (2.00)
July		2.80 (2.80)	4.30 (3.70)	3.60 (3.70)	5.30 (3.70)
August		2.50 (2.20)	3.30 (2.00)	1.80 (1.50)	2.70 (1.50)
Totals	**Water loss**				
	cms.	9.20	12.00	9.10	15.40
	ins. (approx)	(3.70)	(4.80)	(3.60)	(6.20)
	Precipitation				
	cms.	6.70	7.60	7.20	7.20
	ins. (approx.)	(2.70)	(3.00)	(2.90)	(2.90)

Source: Estimated by author based primarily on Appendix D data.

weight of the evaporimeters this would indicate that the cans prevented the inflow of soil moisture supplies from below and laterally. It may be assumed that natural evapotranspiration at the site would be at least as much as that calculated for the evaporimeters. However no impediment to additional soil moisture supplies would exist under natural conditions. The amount by which the inches change in soil moisture exceeded the inches moisture change in the weight of the evaporimeters may be taken as an indicator of the extra evaporation likely to have occurred naturally.

It is reasonable to expect more additional evaporation to have occurred naturally since the less moist environment of the evaporimeters would lower their evaporability somewhat. It is not possible to judge the precise significance of this factor. Baier (1967, p 171-176) has indicated the existence of several different schools of thought concerning the relationship between the availability of soil moisture and actual evapotranspiration rates. Most of this debate has been related to agricultural crops. The nature of the relationship for peat/moss conditions does not appear to have been studied in any detail. As will be shown below the minimum necessary adjustment factor calculated for a one month period at one site was +60 percent. This corresponds to a 20 percent change in volumetric soil moisture for the evaporimeter compared with only a 3 percent change under natural conditions during the same period. In this case there would appear to be good justification for expecting significantly reduced evaporation rates from the

TABLE 5-19

Monthly and Seasonal Adjustment Factors For The Evaporimeter Data

(percentages)

	YT2	YT4	YT6M	YT6S
June	+50	+60	+35	--
July	- 9	+15	+35	+3
August	NIL	+30	+ 5	NIL
SEASON	+15	+40	+25	NIL

Source: Calculated by the author. See text for explanation.

evaporimeters by comparison with the natural conditions. It is therefore likely that the correct adjustment factor would be somewhat greater than +60 percent. This is the extreme case however and generally the minimum necessary adjustment factors were in the range of -8 to +40 percent and the actual changes in moisture are quite small by comparison with the absolute amounts involved.

In one case the change in soil moisture was greater than the change in weight of the evaporimeters indicating that the cans prevented drainage of surplus moisture to the surrounding peat. In this case the difference was subtracted from the evaporation calculated from the evaporimeters.

Monthly and seasonal adjustment factors calculated from a comparison of changes in soil moisture and the change in weight of the evaporimeters are presented in Table 5-19. The monthly values may be considered approximate in view of the degree of averaging incorporated into the calculations. The seasonal adjustment factors can probably be considered as being adequately representative of necessary changes. The results presented in table 5-19 indicate the strong influence of the shallow evaporimeters in impeding horizontal and vertical soil moisture flow. Deeper and larger cans would help to obviate this influence. Impedance of soil moisture inflow appears to have been the major effect, particularly during June. In this month the sites tend to be especially moist because of surplus snowmelt runoff detained in small depressions. The melting out of the active layer or seasonally frozen zone also contributes moisture and since the unfrozen soil zone is still relatively shallow, drainage of this excess moisture is impeded. The influence of the evaporimeters in preventing drainage is indicated for July at YT2. This effect is understandable since the seasonally frozen zone at this site completely thaws out during July, considerably improving natural drainage. The effect continues during August to such an extent that the evaporimeters simulate natural conditions satisfactorily. At YT6S evaporation demands are significantly reduced because of the well-shaded nature of the site. The soil moisture reserves contained in the evaporimeters are adequate to meet these low demands.

/ The results of applying adjustment factors to the initial evaporimeter results, (Table 5-18), are presented in Table 5-20. It is probable that the values are slightly low since the adjustment factors are minimal and do not take account of extra evaporation which may have occurred had the evaporimeters maintained natural moisture conditions. An upward adjustment for undermeasurement of actual precipitation by the rain guages at the sites could also be inserted. It is felt that the combined effect of introducing these adjustments would raise the summer totals of water loss by not more than about 10 percent.

The results obtained from the three main summer months have been extended by estimating early and late season evaporation during 1972. In May the snow did not clear from most sites until about the middle of the month. The second half of May was a fairly warm period and the sites would have

TABLE 5-20

ADJUSTED MONTHLY ESTIMATES OF WATER LOSS FROM THE EVAPORIMETERS AT YT2, YT4, YT6M, YT6S. May-Sept. 1972

(cms.)

	YT2	YT4	YT6S	YT6M
May (estimate)	2.50	2.50	2.00	2.50
June	5.85	8.80	3.70	10.00
July	2.50	4.00	3.65	7.15
August	2.50	4.30	1.80	2.85
September (est.)	1.50	1.50	1.00	1.50
June-August total	10.85	17.10	9.15	20.00
	(4.30 ins)	(6.70 ins)	(3.60 ins)	(7.90 ins)
Estimated Summer total (approx.)	14.85	21.10	12.15	24.00
	(5.80 ins)	(8.30 ins)	(4.80 ins)	(9.50 ins)

Source: Estimations by the author based primarily on field data and calculations.

been very moist as a result of the snowmelt. Evaporation trends during June reveal relatively high rates particularly early in the month. If projected into May and dampened slightly owing to lower net radiation input and the need to warm the sites then evaporation values of about 2 to 2.5 cms. may be estimated.

Because of the relatively early arrival of winter conditions in mid-September it is improbable that late season evaporation was very significant during 1972. The evaporimeters were weighed in early October but it was not possible to safely estimate evaporation since the previous late August weighing because of the difficulties in accurately gauging the combined snow and rain precipitation at the sites during the period. There was some indication however that evaporation was limited to a range of about 1.0 to 1.5 cms. for the month of September.

The results presented in Table 5-20 indicate that despite the inclusion of the adjustment factors the levels of evaporation still appear to be very low by comparison with that which may be anticipated from available net radiation and soil moisture considerations. This conclusion would be only slightly weakened if the additional adjustments indicated above were incorporated. Alternatively the conclusion could be strengthened when it is remembered that conditions during the summer of 1972 favoured above average evapotranspiration.

The highest rates of evapotranspiration occurred at YT6M. Moisture conditions at this site were above field capacity for most of the summer. The relatively low albedo favours a high level of net radiation input. The dominant moss cover, Polytrichum commune, has some vascular tissue (central woody axis) making it a relatively efficient medium for moving moisture from the basal layers to the surface. It appears to have conductive strands like a few other mosses, but it is not proven that such strands actually conduct water. (Rouse, reporting for K.A. Kershaw, Personal Communication, June 4th, 1973). The comparatively low evaporation rate during August is probably related to the soil moisture falling below field capacity and the relatively low net radiation input. It must be stressed that YT6M is not truly representative of the ground vegetation of the spruce woodland environment because of disturbance effects. A truer representation of evaporation from the ground surface of the bog is provided by YT6S, which by contrast manifests the lowest rates of evaporation for the sites investigated. The major influence appears to be that of shading by the relatively dense vegetation considerably reducing net radiation input at the ground surface. The moss cover is still dominated by Polytrichum commune, though some non-transpiring lichen, Cladonia chlorophaea is locally abundant. Moisture conditions probably remain at or near field capacity for most of the summer.

At YT2 relatively low evaporation rates were observed. This site was well-drained by comparison with the others, limiting moisture availability to an important degree. The dominant moss cover appeared to be Bryums. These mosses are very densely tufted and may be considered as relatively inefficient conductors of moisture upwards from the moist basal layers.

A Bryum was also present as the dominant moss understory at the sedge meadow site. In effect the moss cover accounts for about 80 percent of the ground surface, the stems of the Carex making up the remainder. Transpiration through the Carex may have been an important component of the total evapotranspiration for the site. It is possible however that the relatively cold, poorly aerated and nutrient deficient nature of the sedge meadow soil inhibited transpiration as Addison (1970) has indicated for the sedge in his Devon Island studies. The relatively high albedo of the Carex may also have contributed to lower the available net radiation at the site. It would appear that some or all of these possibilities must be invoked to explain the low levels of evapotranspiration at a site where soil moisture conditions are close to a very high field capacity value.

Some indication of the inefficiency of evaporation through these covers may be provided by calculating the crude monthly Bowen ratios indicated by the results obtained. This was accomplished by assuming that the soil heat flux at each site, in each month, was 15 percent of net radiation. Net radiation data were based on the pattern measured at YT2. At YT4 the net radiation input was assumed to be 5 percent greater than at YT2. At YT6M the net radiation was assumed to be 10 percent greater than at YT2. The results are presented in Table 5-21.

TABLE 5-21

Approximate Monthly Bowen ratios
at YT2, YT4, YT6M. Summer 1972

Assumptions: Ground heat flux = 15 percent Rn

RnYT4 = RnYT2 + 5 percent RnYT2

RnYT6M = RnYT2 + 10 percent RnYT2

	YT2	YT4	YT6M
JUNE	1.05	0.45	0.34
JULY	3.60	2.00	0.77
AUGUST	1.60	0.47	1.51

Source: Calculations by the author. See text for explanation.
See Appendix C for information on Bowen Ratio.

In temperate latitudes the Bowen ratio is typically about 0.10 over moist soils but is much larger where soils are well below field capacity (Thornthwaite and Hare, 1965, p.166). The crude Bowen ratios determined in the present study are therefore surprisingly high when it is remembered that at all sites the soil moisture conditions were at or very near field capacity during the whole summer. If the available literature is interpreted in Bowen ratio terms however the present results are seen to be in order.

The Point Barrow studies conducted by Mather and Thornthwaite (1956; 1958) indicated an annual Bowen ratio of about 1.4. (p 87). The monthly ratios obtained for June, July and August in 1958 were respectively 2.0, 2.9 and 1.9 (p 84). In his interpretation of the results of the evapotranspirometer studies made by Nebiker and Orvig (1958), Hare (1965, p 10) infers a summer season Bowen ratio of about 2.0 or greater to be applicable for the lichen surfaces monitored.

The P.E. studies made by Brown (1965) at Norman Wells indicated that the indigenous sedge, moss and lichen vegetation evapotranspired considerably less moisture than the Kentucky bluegrass. Evapotranspiration through the latter cover may be assumed to have manifested very efficient use of the available net radiation and the Bowen ratio would be close to zero. Probable Bowen ratios for the other indigenous covers can be estimated at about 1.0. It is likely that this value would be higher under natural conditions, when *no* irrigation is applied to maintain P.E. conditions, or in areas with less natural summer precipitation than is the case at Norman Wells. Also the moss cover present at the two moss sites consisted of Sphagnum which, by moss standards, is a relatively efficient conductor of moisture from moist basal layers to the surface. This is accomplished through capillary rise, or in the manner of a wick.

Similarly high Bowen ratios have been indicated in other studies in Arctic regions. Ahrnsbrak (1968), in Central Keewatin, indicated very high Bowen ratios for short periods during the summer. At Axel Heiberg Ohmura (1972) found Bowen ratios ranging from 1.2 to 6.0, and averaging 2.5, for fortnightly periods during the summer of 1970. Addison (1973) at Devon Island found a similar range for a series of microsites averaged over two clear July days in 1972.

Heat budget studies conducted by W. R. Rouse and associates in the Hudson Bay Lowlands have indicated relatively low Bowen ratios for surfaces in this region compared with values discussed above. Average ratios for short periods during July were 0.61 for a lichen mat and 0.71 for an old burned lichen cover (Rouse and Kershaw, 1971, p299-301). At a lichen dominated raised beach ridge during July and August 1971 the mean value for the Bowen ratio was 0.73 (Rouse and Stewart 1972, p1064,1067). For one clear July day in 1972 the ratio was 1.29 for an upland, lichen dominated, ridge site and 0.38 for a wet lowland swamp, and 0.38 for a lake (Rouse, 1973). These values are still high enough to indicate that evaporation is not particularly efficient for surfaces in this region in terms of the available radiant energy. Although there is some comparability of the values with those determined for the present study (Table 5-21), the Hudson Bay Lowland values still appear to be generally lower except in the case of the very wet sites.

It is difficult to explain the apparent discrepancy between the Bowen ratios for the Hudson Bay Lowlands and those for other subarctic and arctic regions. The sites investigated lie in a region straddling the open woodland/forest-tundra transition of the Boreal Forest (Hare and Ritchie, 1972, p 338) and may be considered subarctic in nature. Distinctive factors about the sites are the close proximity of Hudson Bay and the relatively low latitude location. Rouse and Stewart (1972) noted that on cold days with the wind blowing off Hudson Bay the Bowen ratios were substantially higher than on warm days with winds from the south. The mean Bowen ratio for warm days was about 0.55 whilst for cold days the mean ratio was almost 1.00. This information aids an explanation of the high Bowen ratios found in Arctic regions since it may be anticipated that summer temperatures in the latter region will be as low or lower than those found in the Hudson Bay Lowlands. However relatively high summer temperatures, particularly maxima, may occur in subarctic areas influenced by a continental rather than maritime climate. This would be the case in the present study area and relatively low Bowen ratios would be expected following Rouse and Stewart's findings. It may also be noted however that relatively low precipitation is also associated with continental climates. Where moisture supply is limited sensible heat flux and hence Bowen ratios will assume greater proportions. Annual precipitation at Yellowknife is about half of that found in the Hudson Bay Lowlands. It would therefore appear logical to expect higher Bowen ratios in the present study area than is the case in the Hudson Bay Lowlands.

The literature on Bowen ratios determined in northern studies seems to indicate a broad trend of increasing Bowen ratios from south to north. Although this trend is probably controlled by decreasing temperatures and precipitation northward it indicates the potential significance of latitudinal position on the compartmentalisation of net radiation flux. The clear July day net radiation values presented by Rouse (1973) are significantly higher than those observed for clear July days in the present study area in 1972 even after allowing for albedo influences. As the net radiation total decreases it appears that Bowen ratios tend to increase. Although the full details of this relationship are not presently well understood it may be used here as an additional justification for expecting higher summertime Bowen ratios in the present study area by comparison with the Hudson Bay Lowlands.

Some insight into the unusual pattern of water loss through moss and lichen covers may be gained by viewing such covers as lilliputian forests, (Bland, 1971). The uppermost surface of the cover intercepts much of the radiation input, reducing considerably the effective radiation reaching the major diffusion exchange surface. This process is hindered by the negligible wind speeds found at these levels in the forest; also the high humidities maintained just above the exchange layer further restricts water loss. These restrictions cannot be bypassed by the efficient system of pipes, (roots, trunks, branches, leaves), for moving moisture found in conventional forests. The moss and lichen covers are non-vascular with only rudimentary capacity for moving moisture upwards.

During periods following rain events, or the snowmelt period, the cover by virtue of its great hygroscopicity intercepts almost all the rain precipitation falling on it except in the case of particularly intense storms when some infiltration may occur. This moisture held in the living plant cover evaporates rapidly, (especially in the case of mosses, but less rapidly in the case of lichens), since the plants possess no mechanism by which evaporation can be controlled as is the case of case with vascular plants (ie. stomates). This point explains the relatively high rates of evaporation, by comparison with the present study, mentioned in the literature survey. Generally the high rates may be associated with high summer precipitation levels. The covers soak up the precipitation and hold it at or near the surface until drying conditions occur. Where summer precipitation levels are relatively low, as in the case of the present study area,the level and pattern of evaporation from moss or lichen covers is closely governed by the precipitation events, especially once the snowmelt runoff influence has passed.

The view is taken in this thesis that transpiration through the vascular flora in the study area is well below normal levels found in more southerly latitudes. This view is based on the literature review presented earlier. It is accepted that there may be significant differences in transpiration rate from individual species making up a plant association. In this study these differences have been generalised out in the overall analysis by considering surfaces (broadly defined) and a seasonal time-scale (rather than individual days or hours). Some support for the usefulness of this admittedly general approach may be derived from the work of Penman (1963). It is felt that his strong emphasis on weather factors over plant factors as primary determinants of evapotranspiration levels may have to be modified in northern areas where permafrost influences appear to alter the functioning of vascular flora and limit transpiration.

The above view may appear extreme in the light of the findings by Belotserkovskaya, Largin and Romanov (1969). Transpiration from non-dominant vascular bog flora was deemed to have been responsible for over 80 percent of the total evapotranspiration. This is acceptable since growth conditions were relatively favourable. Volumetric moisture contents were in the range of 30-76 percent indicating better aeration of the soils. It is probable that temperatures in the soil were also greater than is the case for permafrost-underlain peatland in the present study area. It may be noted that despite these favourable factors total evapotranspiration at the Russian site was in a similar range to that found at YT6M. This similarity is probably due to the lower ground water levels in the Russian study reducing evapotranspiration potential but being compensated for by the exclusively non-vascular and relatively inefficient evaporating moss surface at YT6M. The results from YT4 are also close to the Russian findings. Again this is likely to be due to the relatively low ground water levels in the Russian study rather than significant transpiration through the *Carex* at YT4. The explanation by Addison (1973) for the low transpiration rates through the sedge at Devon Island is thought to apply to the sedge at YT4. This involves a combination of shallow vapour pressure gradients at the ground surface and limited root permeability induced by the influence of the cold nature of the soil inhibiting essential biochemical reactions.

EQUILIBRIUM MODEL

The equilibrium evaporation model suggested by Rouse & Stewart (1972) was applied to data derived during intensive research periods for which evapotranspiration estimates were also available for the evaporimeters. The results of the comparisons are presented in Table 5-22. It should be noted that the initial evaporimeter estimate was multiplied by the seasonal adjustment factor calculated for each site.

The best comparisons between the model estimates and calculated values were obtained at YT4. It is thought that moderately dry conditions prevailed at this site owing to rain events during or just prior to the study periods. At YT2 and

TABLE 5-22. SELECTED APPLICATIONS OF THE ROUSE-STEWART MODEL AT THE STUDY SITES

	Date	Time Period. (M.D.T.)	Net Radiation (cal. sq.cm. min)	Mean Temperature (° c)	Bowen Ratio	Evapotranspiration Model	Evapotranspiration Evapori-meters
						Totals (cms.)	
YT2	11 July	04-08	11·3	7·56	1·10	.24	.12
		08-13	102·0	16·56	0·76		
		13-18	147·1	20·50	0·67		
		18-22	32·9	20·28	0·61		
	24 July	07-12	72·7	21·28	0·63	.24	.06
		12-16	112·6	26·06	0·53		
		16-21	67·0	26·06	0·50		
	9 Aug.	08-12	23·0	15·78	0·81	.06	.07
		12-17	45·1	15·67	0·80		
		17-21	9·1	14·06	0·83		
YT 4	18 July	05-09	40·2	11·78	0·90	.29	.31
		09-13	117·6	18·22	0·71		
		13-17	126·6	22·72	0·55		
		17-21	45·0	21·84	0·54		
	4 Aug.	06-11	41·0	8·28	1·01	.16	.20
		11-15	102·0	12·50	0·87		
		15-20	66·0	14·00	0·82		
	22 Aug.	06-10	31·8	15·72	0·72	.20	.20
		10-15	120·0	26·22	0·46		
		15-19	57·0	28·83	0·40		
YT 6	16 July	06-11	78·0	10·72	0·76	.24	.25
		11-16	178·8	15·50	0·79		
		16-21	67·2	15·39	0·76		
	26 July	07-11	72·5	23·11	0·56	.30	.18
		11-16	161·5	27·06	0·46		
		16-20	81·0	23·89	0·49		
	13 Aug	07-11	49·0	16·61	0·72	.21	.10
		11-16	146·6	21·33	0·62		
		16-20	46·7	19·50	0·67		

Source: Rouse and Stewart (1972),

fieldwork and calculations by the author.

YT6M the best comparisons were also obtained under similar conditions. The poorest comparisons may be associated with very sunny days preceded by a prolonged dry period. It is felt that the latter type of situation was particularly common in the study area and tends to be slightly underrepresented by the comparisons shown in table 5-22. It may also be noted that the good comparisons tend to be associated with relatively cooler temperature conditions reflecting the strong control of air temperature in the model's make-up.

The Bowen ratios indicated by applying the model tend to be greater than the range set for accuracy of the model estimates to be within 10 percent of actual values. This casts doubt on the accuracy of the model estimates, and, coupled with the crude approximations incorporated into the model data and the evaporimeter data, in turn casts doubt on the validity of making the comparisons attempted here.

R. G. Wilson, (Personal Communication, April 27th, 1973), has indicated the admissability of using a version of the equilibrium model to estimate monthly equilibrium evaporation where: $LE = \frac{S}{S + \gamma}(Rn - G)$ where S is the slope of the saturation vapour pressure versus temperature curve at the mean of the wet bulb temperatures at the surface and in the overlying air, γ is the psychrometric constant. The term $(\frac{S}{S+\gamma})$ determines the proportion of the available energy used for equilibrium evapotranspiration and it varies with air temperature. Wilson and Rouse (1972, p 441), have shown this relationship graphically and it is possible to extrapolate

TABLE 5-23

Comparison of Monthly Evapotranspiration
Estimates by the Model and from the Evaporimeters
Summer 1972 (cms.)

		YT2	YT4	YT6M
JUNE	Model	7.0	7.3	7.7
	Evaporimeters	5.9	8.8	10.00
JULY	Model	7.1	7.5	8.0
	Evaporimeters	2.5	4.0	7.2
AUGUST	Model	6.6	3.8	4.4
	Evaporimeters	2.5	4.3	2.9
TOTALS	Model	20.7	18.6	20.1
	Evaporimeters	10.9	17.1	20.1

Source: Calculations by the author. See text for explanation.

values of the term for different air temperatures.

An attempt was made to apply this equation to derive some estimates of monthly evaporation at YT2, YT4 and YT6M. The net radiation data for YT2 was used to estimate net radiation at YT4 and YT2. For YT4 Rn. YT2 was increased by 5 percent and for YT6M Rn. YT2 was increased by 10 percent. Soil heat flux was assumed constant at 15 percent of net radiation. The temperature value used was the mean monthly air temperature for Yellowknife Airport. $(\frac{S}{S + \delta})$ was abstracted from the graph prepared by Wilson and Rouse (1972, p 441). The results are presented in Table 5-23 together with the adjusted monthly evaporimeter estimates.

The monthly comparisons tend to be generally poor and inconsistent. Particularly good comparisons may be noted for YT6M however. This was the wettest site visited and the relative success at the model indicates that 'moderately dry' would be a valid description of conditions. If this is the case almost all other surfaces in the study area can be interpreted as 'very dry'. Particularly poor comparisons were obtained at YT2, the 'driest' site investigated despite volumetric moisture contents in the surface organic layer ranging between 70 and 80 percent.

The attempts to apply the equilibrium model to data obtained in the present study have not been entirely satisfactory. The quality of the initial data is suspect casting doubt on the validity of the comparisons made. It is possible that the model does not apply so well in the present study area because of climate and latitudinal differences. The summer climate in the study area is characterised by brief wet spells and prolonged dry periods. When moistened by rain the moss covers lose that moisture very rapidly by evaporation since they have no mechanism by which to slow the process. The process only proceeds however until the surface layers of the moss cover are dried out and evaporation from the moist soil layers, through the moss, is inhibited by a strong mulching effect. Thus it may be appreciated that "moderately dry" periods will not be a very common occurrence. Very dry surface conditions are probably the rule in the present study area, 'surface' here referring to the top one to two inches of vegetation cover.

The model appears to be insensitive to differences in degree of diffusive resistence between different covers. For instance, certain moss species such as *Sphagnum* appear to have a lower diffusive resistance than others such as *Bryums*. There are apparently also differences in diffusive resistance between different lichens (eg. Addison, 1973). An additional complexity is variations in soil moisture content. It would appear that perhaps the model is too great an oversimplification of the evapotranspiration process through northern vegetation covers. Despite claim for high-latitude application potential the model was developed and tested in relatively low latitudes. It is possible that the compartmentalization of energy varies by latitude and that regression coefficients obtained at the test site do not apply at other latitudes unless modified.

CHAPTER VI

THE WATER BALANCE OF BAKER CREEK

INTRODUCTION

In this chapter an attempt is made to provide a synthesis of the water balance of Baker Creek basin based on the results of the literature and field research reported earlier and an evaluation of the results gained by three separate approaches to determining the annual water balance of the basin:

(a) the Thornthwaite climatic water balance model;

(b) climatonomic analysis;

(c) semi-empirical studies.

In effect this chapter is a means of summarising and applying the total body of research that has been conducted in the course of the present study.

THORNTHWAITE CLIMATIC WATER BALANCE

The Thornthwaite P.E. concept and water balance model has been widely used to obtain an insight into the water balance patterns in a particular area. It is considered here for its ability to produce working estimates of actual evapotranspiration for the basin. The estimates are then subjected to an evaluation based on results from the present research.

Kakela (1969, 1973) applied the concept and model to data from the present study region for the period 1942-1967. Kakela's work has been updated in the present study by applying the model to data for the water years 1968-1972. The results are presented in Table 6-1. The 25 year mean obtained by Kakela (1969, p. 216) has been included for comparative purposes. It may be noted that average precipitation during the period 1968-1972 was over half an inch greater than the 25 year mean, and P.E. was on average almost one inch less. Calculated surplus or runoff is very similar for both periods however. A one inch soil moisture storage capacity was used in both cases and P.E. was calculated using the 50°N correction factor, this being the highest latitude covered by the Thornthwaite computation tables (Thornthwaite and Mather 1955, 1957).

The relatively large discrepancy between the average P.E. values for the two periods is surprising. In total the differences between the averages for the two periods were that on average during the period 1968-1972 the summers have tended to be wetter on average with the opposite being the case for winter snow precipitation. The wetness of the recent period summers has been concentrated markedly during the months of August and September. (See Figure 5-5). It is likely that the cool, cloudy conditions associated with these abnormally high precipitation levels were in large part responsible for the lower P.E. rates during this period. It may be noted that average actual evapotranspiration (A.E.) was greater during the recent period by an amount closely corresponding to the difference in average precipitation for the two

TABLE 6-1

THORNTHWAITE WATER BALANCE FOR STUDY AREA

WATER YEARS 1968-1972, YELLOWKNIFE AIRPORT DATA

Water Year	Precipitation	=	(P.E. - Deficit)	+	Surplus ± Storage charge
1968	11.39	=	(14.75 - 9.64)	+	5.77 + 0.51
1969	11.93	=	(14.96 - 6.83)	+	4.31 - 0.51
1970	9.86	=	(15.16 - 7.50)	+	2.20 ± 0
1971	9.00	=	(16.05 - 9.84)	+	2.79 ± 0
1972	9.65	=	(15.41 - 10.01)	+	3.70 + 0.55
Mean 1968 - 1972	10.36	=	(15.26 - 8.76)	+	3.75 + 0.11
Mean 1942 - 1967 (Kakela, 1969 p. 216)	9.74	=	(16.19 - 10.28)	+	3.81 + 0.01

N.B. A one inch soil mositure storage capacity was used

Sources: Calculations by the author; Thornthwaite and Mather (1955, 1957).

periods. This is to be expected in regions of low precipitation relative to P.E., especially where significant detention storage exists. It may be remembered that the Point Barrow studies indicated a virtual balance between precipitation and evapotranspiration during the three summer field seasons (Mather and Thornthwaite, 1958, p. 93).

The relatively dry winters during the recent period raise the possibility that runoff may have been reduced during this period but this influence seems to have been balanced out by detention storage being maintained at high enough levels to result in average runoff values prevailing. It may be noted that there was only a very slight difference between the estimated level of Pocket Lake in December 1968 and December 1972. (Figure 5-8; Table 5-7). These broad considerations drawn from the Thornthwaite model also tend to substantiate the earlier assertion that the available Baker Creek runoff record, although short, yields good values for average conditions. In view of the emphasis placed on use of this short record this indication of support is valuable.

It may be noted that the calculated runoff by the Thornthwaite method, ($\approx$ 3.80 inches), is considerably higher than the average measured runoff for Baker Creek from 1968-1972, (1.43 inches), Table 5-6. Actual evapotranspiration by the method is in the range 6.0 to 6.5 inches. It is probable that this range would be greater if the water balance was adjusted to take account of the low measured runoff. This adjustment could be most easily accomplished by amending the soil moisture storage assumption in an upward direction to approximately 3.30 inches on an average basis.*

If it were assumed that the Thornthwaite approach did give accurate estimates of runoff in this region, one way that runoff values could be made to approximate measured runoff would be for soil moisture storage capacities to change from year to year. From Table 5-6 it may be deduced that in individual years the storage capacity required to balance the calculated and measured runoff would range from approximately 1.80 inches to 5.25 inches. This range is quite conservative by comparison with a similarly derived range, (20.3-0.3 cms.), estimated by Kakela, 1973, p. 176), for 1942-1967 data. It encourages the belief that storage capacity inputs to the P.E. model may be indexed by variations in the lake level. However an attempt to relate Pocket Lake level patterns, (Table 5-7), and the soil moisture storage assumption necessary for a balance met with no success.

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* It is possible that a more valid approach to this form of averaging soil moisture storage capacity is to calculate the Thornthwaite water balance for a number of storage values (approximately) corresponding to those for a number of different site types which are prominent in the basin. If desired, area weighting procedures would then permit an estimation of the average basin water balance.

It does appear however that the one inch soil moisture storage capacity originally employed by Kakela (1969, 1973) and in the present study is too low and that a better average value would be about 3.30 inches. This of course would raise the A.E. predicted by the model to between nine and ten inches. It would be desirable to use the model to derive separate evapotranspiration estimates for land and water surfaces. The long term mean P.E. calculated by Kakela is very close to the evaporating power of the net radiation available after snowmelt energy has been allowed for (16 inches) (Table 3-15) The 1968-1972 mean P.E. is very close to the mean annual small lake evaporation estimated for the study area on the basis of Pocket Lake lake level records and other sources (Tables 5-11, 5-12). These correspondences indicate support for a free water evaporation in the study region of 15 to 16 inches. Assuming that lakes account for 25 per cent of the drainage basin and that a 3.30 inches storage capacity is employed the above range estimate for mean lake evaporation would result in a mean land A.E. value ranging between 7.3 to 7.7 inches. An indication of how this value is spread over a variety of surfaces can be guaged by estimating the water balance using different storage values for different site patterns as Kakela (1969) has done in a speculative way. This latter work forms a comparative base for the final section of this chapter.

The indications of relatively high Bowen ratios over surfaces in the study area cast doubt on the validity of the Thornthwaite procedure for computing P.E. for high latitude regions. Hare (1972) has indicated that the method implicitly assumes that Bowen ratios are zero or very low over the idealized evaporating surface. It may therefore be anticipated that the procedure will be refined or replaced as a result of research into northern radiation balance patterns. Kakela (1973) has indicated (for the Yellowknife and Snare river basins) that this present shortcoming of the procedure is not so serious as the questionable book-keeping assumptions (especially those relating to storage factors), that it involves. To a large extent these assumptions are removed when considering daily water balances of various sites for individual years rather than the mean annual basin balance based on monthly averages. The former ideal is difficult to realise however, partly because of the time-consuming nature of the calculation process and partly because of the lack of detailed knowledge on the nature of the interrelationships of precipitation, evapotranspiration, runoff and storage change at each site.

Because of the relatively high Bowen ratios found for northern regions it is acknowledged that P.E. will be equivalent to a smaller portion of the net radiation in the study area latitude than further south. The present research indicates however that large deficits may still be anticipated over vegetated land surfaces in the Subarctic, despite the presence of soil moisture levels at or near field capacity one or two inches below the top of the vegetation cover. At some sites, such as the muskeg, one of the major reasons for the low A.E. is the relatively good drainage reducing water supply. This is generally the major reason, for P.E. not being realised at most mid-latitude sites. The exact nature of the moisture comparison should be stressed however since the muskeg is relatively dry in relation to the bog and sedge meadow but it would probably be termed very moist in relation to most mid-latitude soils. An additional factor limiting P.E. at the muskeg site is the extensive moss cover. Such a cover is also the dominant ground story vegetation at the meadow and bog sites. Mosses (and lichens) are non-transpiring plants and can also function as effective mulches thus endowing these covers with high diffusion resistance properties when dry.

The effect of the hygroscopic properties of mosses is particularly noticeable at the muskeg site. Especially during the later summer months, after snowmelt influences have diminished, almost 100 per cent of A.E. is accounted for by precipitation receipts held in the moss canopy and lost easily when drying conditions resume. P.E. might be attained briefly only during such periods partly since mulching effects are minimal and partly because the mosses have no well-developed physiological mechanisms for controlling moisture loss.

At the much less well-drained permafrost sites the influence of upward supply of moisture to the surface region from lower soil layers assumes a greater importance by comparison with precipitation in explaining the evaporation totals measured or estimated for the sites. Detained snow melt surplus appears to be primarily responsible for the high totals however. In addition the force of moisture supply from below may help to exploit any moisture conducting properties possessed by the mosses. Operation of this approximate substitute for the transpiration process, although not proved to exist, appears to be favoured for certain species such as Sphagnum and Polytrichum commune, and less favoured for others such as the Bryums.

Evaporation from the shaded sub-site at the spruce bog was the least of the four sites investigated. It is possible that if transpiration through the shading trees is taken into account that the overall evapotranspiration for the site would be much greater than for the ground moss/lichen cover alone. No field research could be undertaken to substantiate this hypothesis, however, interpretation of results from the literature survey in conjunction with an evaluation of local conditions tend to indicate that high transpiration losses may not be anticipated for the higher order vascular flora, especially over permafrost sites. The main inhibiting factor appears to be a low degree of root permeability, precipitated by the cold soils, which reduces the efficacy of the plant system to maintain a moisture supply to meet demands generated by meteorological conditions above the surface. This reasoning appears to explain the apparently limited transpiration through the Carex at the sedge site. The explanation manifests another significant restriction on the realisation of optimal P.E. conditions in northern areas.

A ramification of the occurrence of relatively low A.E. from vegetated surfaces is the increase in the importance of the sensible heat flux component in the radiation balance of the present study area. When considered in advection terms this may well have the effect of significantly supplementing the net radiation energy available over the bedrock-dominated sites and the free water surfaces. The influence is undoubtedly greater for the latter category where moisture supply is unlimited. Indeed it may be anticipated that the bedrock surfaces will supplement the advective energy supply to the water surfaces from the vegetated surfaces. As a result significantly higher P.E. values may be realised over the water surfaces than is indicated solely by net radiation considerations. This pattern will be emphasised in the cases of the smaller free water bodies where heat storage influences are negligible. Although earlier discussion in this section indicates mean lake evaporation in the study region to be 15 to 16 inches, (i.e. at or slightly below the amount that may be expected on net radiation grounds), later evidence which is considered superior because of greater consideration of local advection influences, indicates that evaporation from all free water surfaces in the basin averages 18 inches or greater.

CLIMATONOMIC ANALYSIS

The apparent potential of the function, derived by Lettau (1969) relating the climatonomic runoff and dryness ratios as a means of generating and testing hypotheses concerning perfect water balance situations has already been noted. It is possible that the function fails in Canada but it appears to predict well the water balance for Canadian locations for which good meteorological and hydrological data checks are available. In addition the data produced from the detailed Subarctic water balance study conducted by Findlay (1966, 1969) at Knob Lake resulted in a much improved fit relative to the function for this location. (The improvement was mainly attributable to more precipitation measurements). There are therefore good grounds for assuming that the function may be employed to determine the water balance for a variety of regions in Canada, including the Subarctic zone within which Baker Creek basin is located.

On the basis of results from research into evaporation and evapotranspiration patterns in the basin different possible combinations of precipitation and runoff were used to calculate the runoff and dryness ratios and the fit of the plots to the curve of the function evaluated. This operation could only be performed for climatological mean annual balances since the seasonal balances require a far more sophisticated analysis which, according to Hare and Hay (1971, p. 79) cannot yet be usefully attempted. except for certain localities in southern Canada.

The period for which means were derived is short, consisting of only five years of data (1968-1972). This was conditioned by runoff records only being available for Baker Creek from 1968. It is felt that these runoff records are of much more relevance in determining the water balance in the area covered by Yellowknife Airport meteorological station than is the case for Yellowknife River data. The latter drainage basin is much larger than Baker Creek and extends over an area where bioclimatic gradients change rapidly (Hare and Ritchie, 1972) reducing greatly the likelihood that Yellowknife Airport precipitation data, located just off the southern tip of the basin, is representative of conditions over the whole drainage basin. Therefore although good long-term records exist for the Yellowknife River the short period record for Baker Creek is preferred because of its more valid use with Yellowknife Airport precipitation data. Yellowknife Hydro climatological station, although located close to the upper basin of Baker Creek is not considered as possessing particularly representative precipitation data for this area because of its relatively low elevation the dubious accuracy of the

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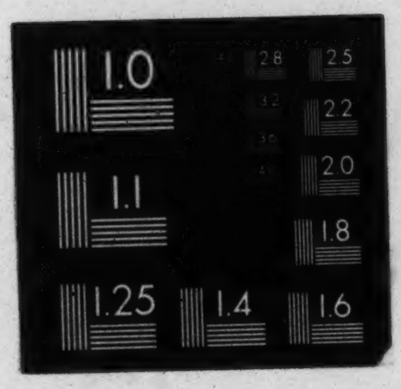

snow precipitation measurements and possibly less careful observation procedures compared with those at the airport.

Apart from precipitation and runoff values an estimate of net radiation is also required to allow calculation of the ratios. Net radiation values are available from Hay (1970). It is felt that previous evaluation of these estimates justifies acceptance of them as a good indicator of mean annual net radiation. The annual value indicated by Hay for Yellowknife is 25.2 Kilolangleys (Kly.) (Hay, 1970, p. 161). Some of this energy is utilised in snowmelt and must be subtracted from the total value. An estimate of this factor is provided by extrapolation from a figure indicating the spatially smoothed field of the energy required to melt the measured fresh snowfall. (Hare and Hay, 1971, p. 92). This was taken to be 1.2 Kly. and the net radiation value used in calculating the dryness ratio was 24.0 Kly. Only the surface energy balance is considered here and it is done so on an annual basis, therefore there is no need to make a similar allowance for the significantly high levels of energy required to thaw out and heat up the surfaces. This is a reversible process at the surface unlike melt energy consumption for which the reverse process (condensation) occurs at cloud level. The potentially great importance in the radiation balance of ground heat flux at the surface over short periods between spring and fall should be borne in mind however. The climatonomic approach followed here involves a regrettable degree of rather unrealistic averaging in time and space which is however necessary in the light of the present unsatisfactory state of the data pertaining to northern water balance patterns.

The base for the precipitation adjustments was taken as the approximate mean for the period 1968-1972 - ten inches. It is assumed that long term storage change is zero. This is substantiated in large part by reference to Table 5-7 and by considering the December 1968 and December 1972 estimates.

Previous to the installation of the Baker Creek Main Stem guage it was only possible to use Yellowknife River data with Yellowknife Airport precipitation to estimate the water balance of the area. Using these values to calculate the ratios results in plot A. which gives a very poor fit. (Figure 3-5).

A. P = 10 inches. N = 3 inches. (E. = 7.0 inches)
C = 0.30 D = 1.53

Kakela (1969) adjusted precipitation data for Yellowknife Airport by increasing snow precipitation by 50 per cent and rain precipitation by 15 per cent on the basis of Russian findings. As a result he obtained a mean precipitation figure of 12.75 inches. This is used for plot B. A much better fit is obtained using this value.

B. P = 12.75 inches. N = 3 inches
(E = 9.75 inches). C = 0.23 D = 1.14

However, the runoff for Baker Creek in individual years and on the average is much below three inches, hence it is very unlikely that this combination is valid for Baker Creek basin. Following a similar process of hypothesis generation, testing and evaluation many possible combinations of P and N were studied. The use of the Baker Creek runoff mean (1.43 inches) immediately produced very good fits necessitating quite precise evaluation of the possible combinations. It is possible that the precision employed in the evaluation is not matched by an equivalent degree of precision in the Lettau function relationship as it applies to Canada generally and to Subarctic locations in particular. This is because Lettau used as a base Budyko's analysis of European drainage basin data, (29 basins), compiled by Wundt (1937). Although evidence has been presented earlier to support the assumption that the curve applies for a variety of areas in Canada with good check data it is possible that incorporation of a representative stratified sample of basins in the less humid and colder regions of Northern Canada would alter the nature of the curve. Only a slight alteration would be required to upset the following precise pattern of reasoning and this should be borne well in mind as probably requiring modification at a future date.

The point has been made that, to a much greater degree than is the case in more humid and warmer southern areas, runoff in dry, cold northern regions is the result of special circumstances such as unusually high precipitation events, bare rock and snow-drifting influences, and unusually low rates of utilization of available energy by evaporative heat flux (Laycock, Personal Communication, August 1973). This point is certainly valid when the climatonomic approach comes to be applied to determining the water balance for individual months in any one year for the present study area. It may be anticipated that numerous modifications will be necessary at this level just as is presently the case in relation to the book-keeping assumptions in the application of the Thornthwaite water balance model in the region. It has been necessary here however to concentrate on average patterns in an effort to obtain a relatively well defined base prior to attempting to remove the time and space averaging assumptions.

The results of calculations using various combinations of P and N are presented in Table 6-2. In order that the closeness of fit of the plot for each combination may be compared an index has been included. This index was obtained by multiplying together the differences between

(a) the plot runoff ratio and the runoff ratio corresponding to the closest part of the curve to the plot, and

(b) the plot dryness ratio and the dryness ratio corresponding to the closest part of the curve to the plot.

From Table 6-2 it may be noted that many of the combinations give relatively good fits with the function,

TABLE 6-2

Computations of Runoff and Dryness Ratios
for Different Combinations
of RUNOFF and PRECIPITATION

PRECIPITATION (inches)	RUNOFF (inches)					
	1.00		1.20		1.43	
	C	D	C	D	C	D
10.00	.10	1.60	.12	1.60	.14	1.60
	(a) 2	(b)14	4	23	6	30
	(a)x(b)28		92		180	
10.50	.09	1.52	.11	1.52	.13	1.53
	0	0	2	10	4	19
	0 (X)		20		76	
11.00	.09	1.45	.10	1.45	.13	1.45
	-1	4	0	0	3	11
	-4		0 (Y)		33	
11.50	.08	1.39	.10	1.39	.12	1.39
	-4	15	-2	5	0	0
	-60		-10		0 (Z)	
12.00	.08	1.33	.10	1.33	.11	1.33
	-5	25	-3	11	-2	7
	-125		-33		-14	
12.50	.08	1.28	.09	1.28	.11	1.28
	-6	30	-5	21	-3	12
	-180		-105		-36	

R = 24.0 Kly. (Hay 1970; Hare and Hay, 1971)

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c = Runoff Ratio   D = Dryness Ratio
(see p.112 for further information on the ratios, or Lettau (1969)
(a) & (b) Refer to text for an explanation
x,y,z These identify combinations referred to later in the text.

Sources: Calculations by the author; Lettau (1969); Hay (1970); Hare and Hay (1971).

similar to that for the Knob Lake improvement. All lay within the box in the inset of Figure 3-5. In three cases however a perfect fit was obtained (X,Y,Z,). The selection of the case which is most probably valid requires close consideration of probable precipitation and runoff patterns. The 10 inch mean precipitation for Yellowknife Airport is almost certainly low. The three perfect fits lie on a diagonal represented by a precipitation range of 10.50 to 11.50 inches. This appears to be a good approximate range for adjusted precipitation, although it requires verification by detailed research in the basin over a period of years.

Considering runoff (N) the mean value obtained for Baker Creek is believed to be a good integrated estimate for the whole basin. There is not a strong case for increasing it due to errors in interpolating records for gaps in the data since during this process a liberal rather than conservative approach was followed. Baker Creek basin is about 15 miles long and the headwaters area lies at about 200-250 feet above the lower basin and Yellowknife Airport. It is conceivable that these contrasts may be sufficient to cause water balance patterns in the lower basin to differ slightly from the upper basin. It is probable that runoff from the upper basin is slightly greater than 1.43 and slightly lower for southern parts of the basin where much of the present and past research in the basin was conducted. A runoff for the southern part of the basin of 1.00 inches appears to involve a much greater contrast than is likely and 1.20 inches is probably a good estimate. This reasoning indicates that the possible mean annual water balance for the lower basin is:

$$\begin{array}{lcccccl} & P & & N & + & E & \\ \text{(inches)} & 11.00 & = & 1.20 & + & 9.80 & \text{(Case Y)} \end{array}$$

Tentative ranges of the parameters for compound assemblages of surface types in various parts of the basin may be set at

	South		North	
P =	10.50	-	11.50	inches
N =	1.20	-	1.60	inches
E =	9.30	-	9.90	inches

SEMI-EMPIRICAL STUDIES

It would be helpful to have some check on the results of the above operation. This may be supplied in part by refining the site water balance equations developed by Kakela (1969) for Pocket Lake sub-basin. Kakela employed the Thornthwaite climatic water balance model to derive the equations and hence included P.E. values. Previous discussion of this chapter has indicated the unsatisfactory level of development of means of estimating P.E. over surfaces in the basin therefore in the present study straightforward mean annual water balances are considered with the assumption that net storage change is zero. An important aspect of Kakela's equations was his use of the adjusted precipitation of 12.75 inches instead of 10 inches. This is considered a large overestimation of the precipitation adjustment necessary. The present research into land evapotranspiration and runoff data in the study area indicates values which lessen the need for such a high adjustment as Kakela has suggested. In the present exercise the site water balance equations have been written for a precipitation base of 11.00 inches divided into about 5.25 for winter and spring periods (Oct. 16-May 15) and 5.75 for summer precipitation (May 16-Oct. 15). Although Kakela considered detailed site seasonal variations in his study only his mean annual site water balance is used in the present study.

In writing the equations particular attention was given to snow drifting changes for each site identified by Kakela (1969). Because of the lower precipitation base adhered to here, Kakela's changes were reduced by a roughly proportionate amount. This procedure permitted an estimate to be made of the net precipitation at each site. In the process to separate this amount into evaporation and runoff certain broad guidelines were followed. For the vegetated sites evaporation was estimated on the basis of results obtained from the evaporimeters (Table 5-20) and a consideration of the drainage properties at each site.

In the case of the bedrock-dominated surfaces several assumptions were made. A consideration of storage capacity values and precipitation data indicated that runoff from these surfaces may not have been as great as Kakela indicated. Although storage capacities appear low for bedrock dominated surfaces the nature of the precipitation events are such that sufficient capacity exists on most precipitation days to hold most if not all the precipitation. (This argument is detailed in the lake evaporation results section). Therefore it may be assumed that a large percentage of the summer precipitation is detained and evaporated rather than running off. Snowmelt may be considered as a major precipitation event. The results from the study by Landals (1970, p. 85) were used as guidelines for gauging runoff/storage relationships at the sites.

The results of the exercise are presented in Tables 6-3 and 6-4. Table 6-3 is a description of the study sites mentioned in Table 6-4. The (PE-D) component in Kakela's equations was reduced to one value - actual evapotranspiration. The main differences between Kakela's estimates and those made in the present study are greater evaporation and less runoff from the bedrock-dominated surfaces and vice versa for the vegetative surfaces.

Using Kakela's percentage area weightings for each site the mean water balance for the Pocket Lake sub-basin was obtained. The parameters were observed to correspond closely with the most probable water balance parameters identified by climatonomic analysis. The runoff value given is the runoff from the basin into the lake required to maintain the lake level. It may be noted that the sub-basin evaporation from the present study is very similar

TABLE 6-3

DESCRIPTION OF SITES REFERRED TO IN TABLE 6-4

Site	Description	Surface Retention Capacity (inches)
A	Bare smooth sloping bedrock	0.13
B	Rough flat lying bedrock with lichens	0.50
C	Very rough (plucked) bedrock, slightly weathered bedrock or thin drift covered bedrock; scant vegetation	1.00
D	Thin coarse soil on slope with light tree-bush-moss vegetation	1.50
E	Tree-moss vegetation on flat to gently sloping well-drained soil	2.13
F	Moss-tree vegetation on flat, moderately well-drained soil	3.00
G	Poorly drained moss covered depressions	4.00
(H)	Bog and wet muskeg (does not occur in Pocket Lake basin)	8.00
I	Deep lakes (Pocket Lake)	Unlimited

Source: Kakela (1970, p. 223).

TABLE 6-4

REVISED ANNUAL WATER BALANCES FOR POCKET LAKE SUB-BASIN SITES

Kakela (1969) Values in Brackets

				(inches)				
Site	%Area	P		E		N		S.C.
A	3.8	8.75 (9.50)	=	4.50 (3.58)	+	4.25 (5.92)		-
B	25.9	10.75 (12.25)		6.00 (4.45)		4.75 (7.80)		-
C	11.3	12.25 (14.75)		6.50 (6.95)		5.75 (6.30)		-
D	5.6	12.00 (14.75)		6.50 (8.45)		5.50 (5.17)		-
E	4.8	11.50 (14.25)		7.00 (9.08)		4.50 (3.05)		-
F	8.3	11.25 (13.25)		8.00 (10.20)		3.25 (0)		-
G	4.1	9.75 (10.75)		8.50 (10.75)		1.25 (0)		-
(H)	-	10.75 (12.75)		10.00 (12.75)		0.75 (0)		
I	36.2	10.75 (12.45)	=	15.40 (15.04)	+	- -	-	4.65 (2.59)
Integrated Basin Water Balance	100	10.95 (12.76)	=	9.75 (9.72)	+	2.88 (3.98) (=) 1.20 (3.04)	−	1.68 (0.94)

Source: Estimations by author; Kakela (1969)

N. B. P = Precipitation; E = Evaporation;
N = Runoff; S.C. = Storage Change.

to Kakela's. However a large difference exists between the precipitation and runoff values obtained in either study. It is believed that the values described in the present study are much more representative of water balance patterns in Pocket Lake sub-basin. Kakela (1969) manipulated his equations toward the bases of 3.00 inches runoff (Yellowknife River mean) and 12.75 inches precipitation (his adjusted value). It has been suggested that these are unrealistic values for the water balance in the local region of Yellowknife Airport. It may be noted that the site equations obtained in the present study were obtained only after one slight manipulation of initially derived values.

Kakela (1969, p.68) has observed that there is probably slightly more area of bare bedrock (and open water surface) and slightly less bog and muskeg type soil conditions in Pocket Lake basin than is generally the proportion throughout the Yellowknife area. (The equation for bog and muskeg is included in Table 6-4 although this site type does not occur in Pocket Lake basin). The changes in evapotranspiration and runoff associated with these differences would probably in large part balance one another out although there is some possibility that evapotranspiration would decrease slightly and be balanced by a slight increase in runoff.

The mean land evapotranspiration value for the basin calculated from sites A to G in Table 6-4 is 7 inches. If it is assumed that lakes and other free water bodies constitute 25 per cent of the area of the basin, the integrated basin estimate for free water evaporation is 18 inches. (This is also on the basis of an integrated basin evapotranspiration total of 9.80 inches). The 24 Kilolangleys of net radiation remaining after snowmelt energy has been allowed for amounts to an evaporating power of 16 inches. Thus it would appear that local advection of energy is an important aspect of the heat balance of the free water surfaces in the basin. If it is found that the percentage area occupied by free water surfaces is less than 25 per cent then the importance of the advection influence must rise to explain the pattern realised. The local advection influence appears to be surprisingly strong but it may be noted that the free water evaporation totals indicated lie below that indicated by the Class A pan data (Table 5-5). It is also possible that the mean small lake evaporation estimate for the study area (15.4 inches) derived mainly from Pocket Lake level records (Tables 5-11, 5-12) is conservative on account of the fact that Pocket Lake is relatively deep or relatively well sheltered by comparison with most other areas of lake surface in the basin. It is also possible that riparian runoff into the lake from the rest of Pocket Lake basin was underestimated owing to conservative allowances for runoff from heavy rain events.

From Table 6-4 it may be deduced that bedrock-dominated surfaces (Sites A,B,C.) generate approximately 75 per cent of the runoff into the lake although such surfaces account for just over 40 per cent of the area of the sub-basin. The importance of the bedrock areas as high yield sources is probably enhanced over Baker Creek basin as a whole because of the more frequent occurrence of such surfaces. This reasoning would tend to dampen the yields that may be expected from vegetative depression areas. The high frequency of shallow lake bog and muskeg surfaces in the Duckfish Lake portion of the basin may effectively make this area an internal drainage basin in most years with high evaporation losses. By contrast the adjacent extensive block of almost exclusively bedrock-dominated area in the north-eastern portion of the basin may well account for much of the annual runoff yields measured for the basin as a whole. It is necessary that more research effort be directed to determining the exact patterns for such surfaces in view of their apparent importance in the water balance.

The above attempt to allocate proportions of the average basin runoff over different blocks of surface types in the region should ideally be extended to other blocks in the basin. The drainage area/lake area ratio for various portions of the basin may serve as a suitable index to guide the allocation procedure. In view of the apparent significance of local advection for lake evaporation it may be useful to derive some form of index of the relative distribution of lakes and land areas whereby areas possessing many evenly spaced small lakes would be differentiated from areas dominated by one large lake. It would be particularly useful if the factor of lake depth or volume could be incorporated in either of the indices. Another direction for future research is to determine the temporal variations in the water balance for short periods in individual years. This approach would probably reveal a close relationship between average runoff and relatively few major precipitation events occurring in short periods of certain years with associated conditions precluding significant evapotranspiration.

The above lines of research would contribute greatly to overcoming the major limitation of this synthesising chapter, namely, that of preoccupation with average annual values which tend to be relatively meaningless for most years, especially in regions such as the present study area in which the surpluses are largely the products of well above normal precipitation seasons, months and days and surfaces with relatively low evapotranspiration losses.

CHAPTER VII

SUMMARY, CONCLUSIONS AND SUGGESTIONS FOR FUTURE RESEARCH

SUMMARY

The present study has been an attempt to define more closely the average annual patterns of evaporation and evapotranspiration from surfaces in the Baker Creek Basin, near Yellowknife, N.W.T. This is a small research basin located in the Subarctic climatic zone, the discontinuous permafrost zone and the Canadian Shield physiographic province. Specific emphases in the study have been upon the elucidation of patterns of evaporation and evapotranspiration for the vegetation indigenous to the Subarctic, and of lake evaporation patterns. Perspective on and partial verification of these patterns was obtained by an assessment of the average annual basin-scale water balance.

The physical setting of the study area has been described. A comprehensive literature review has been provided and used extensively in places to fill gaps attendant upon field research instrument problems or to establish trends or influences which potentially affect the present study area.

The major field work portion of the research was conducted at three main sites: muskeg (relatively dry), sedge meadow, and a spruce woodland bog, (exposed and shaded subsites). At these sites net radiation measurements were made or attempted. In addition batteries of evaporimeters were employed to estimate evapotranspiration and associated soil moisture regime studies were conducted. An attempt was made to apply an equilibrium evapotranspiration model to the data. Active layer development was monitored at two sites.

For the lake evaporation studies surface temperature surveys were made over several lakes in the study region with a view to using the data in a Lake Hefner type equation for estimating evaporation. Evaporation from small, intermittent pools was also estimated using monitored changes in level.

Extensive use has been made of recently available hydrological data for the basin. This consisted of lake level records for Pocket Lake, and flow records for its subbasin, available since 1967, and the discharge records for Baker Creek (Main Stem), available since 1968. Meteorological data for Yellowknife Airport have been utilised to provide background climatic information for the various study periods.

An attempt has been made to synthesise a better definition of the average annual water balance of Baker Creek basin by means of three separate approaches:

(a) the Thornthwaite water balance model,

(b) climatonomic analysis, and

(c) semi-empirical research.

By virtue of the latter approach, which integrates all the research results, refined average annual water balance equations have been presented for various surfaces present in the study area.

CONCLUSIONS

Interpretation of the results of field research into evapotranspiration from vegetative surfaces indicated that this process did not function efficiently at the study sites in terms of available radiant energy and soil moisture considerations. This conclusion backs up similar findings noted in the literature review for other northern studies. A general factor explaining this pattern is the significant proportion of radiant energy employed in melting snow, ice and permafrost and raising surface and air temperatures. A more specific explanatory factor was the non-vascular nature and mulching properties of the moss-covered sites. In the case of the sedge, transpiration through these plants was thought to be inhibited by distinctive edaphic factors associated with the permafrost present at the sites. Similar reasoning was employed to suggest that transpiration through the higher order vascular flora would also be inhibited. The precise moisture conditions and the nature of summer precipitation events were also held to be important factors governing the levels of evapotranspiration. The moss covers only rarely appeared to be saturated right to the surface. As a result some mulching effect was always present to shield against what would otherwise have been uncontrolled and significant water losses. At the onset of rain events the upper moss surfaces tended to have good hygroscopic properties. They were thus able to retain much of the incoming moisture which would be evaporated rapidly when drying resumed owing to the plants' lack of physiological controls on evaporation.

Although the bedrock-dominated surfaces in the basin were not intensively investigated conclusions derived from an appreciation of the combined relationship of storage capacities and the nature of the precipitation events indicated that the significance of runoff from these surfaces may have been overestimated in the past and the significance of evaporation from them underestimated. In sum the study indicated that evapotranspiration from land surfaces in the basin averaged about seven inches.

Interpretation of the results of the lake evaporation studies indicates the possibility that the magnitude of the parameter may have been underestimated in the past, possibly due to hitherto unappreciated local advection effects or by previously exaggerated heat storage lag effects and condensation gains. If it is assumed that lakes account for 25 per cent of the area of the basin the free water evaporation over the basin averages eighteen inches. This is two inches greater than the evaporating power of annual net radiation after snowmelt energy has been allowed for. It is four inches greater than the mean small lake evaporation estimates published in the literature. It is however two inches less than the average class 'A' pan evaporation. It therefore appears to be a plausible intermediate value.

Previous water balance estimates for the study area were based on Yellowknife River runoff data but the new data from Baker Creek improved the definition of the water balance considerably, just as better precipitation measurements have allowed improvement of the balance for other northern areas. It became clear that relatively minor upward adjustment of precipitation was required to enable perfection of the water balance for the present study area. Although a case could be made for reducing estimates of evapotranspiration from vegetative surfaces in the basin by comparison with previous estimates, the high level of free water evaporation, in addition to greater evaporation from bedrock surfaces was seen to more than compensate for this reduction. In effect then the finding lessened the likelihood that considerable upward adjustment of precipitation was necessary. It was estimated that the basin precipitation estimate derived from Yellowknife Airport records required only an upward adjustment of about 10 per cent for them to be taken as representative of the basin.

The average annual water balance for Baker Creek basin for the period 1968-1972 may be expressed as:

	Precipitation	=	Runoff	+	Evapotranspiration
(inches)	11.00	=	1.40	+	9.60

P.E. for the study area is probably close to the

16 inches value generally quoted for the area. It may be slightly lower if the relatively high Bowen ratio considerations apply strongly or slightly higher if advection effects are truly significant. In any case it is clear that the average annual basin P.E. will be well above the annual precipitation and the runoff from the basin must be largely the product of exceptional precipitation events, (where snowmelt runoff is also considered as a major precipitation event). This runoff would be greater if depression and lake storage, aiding evapotranspiration, were not such an important feature, in addition to the bedrock-dominated surfaces, of the basin hydrology.

SUGGESTIONS FOR FUTURE RESEARCH

In the course of the present study particular subjects requiring further research in the study area have been indicated. These have been brought together in this section and some additional comments are made concerning some important points of possible use to future researchers.

There is a fundamental need for a meso-scale study of climate in the region, not only as a check on the results indicated here but also to enable the results to be applied over a wider area. The representativeness of Yellowknife Airport data for the rest of the basin and the region must be established. Particular influences worthy of consideration are those of Great Slave Lake and increasing elevation inland. When determined, temperature patterns associated with these influences would be of great utility in estimating resulting evaporation patterns. Knowledge of precipitation patterns, such as snow distribution and thunderstorm effects, conditioned by the influences would be even more useful. The diurnal wind speed pattern appears to be of significance in reducing transpiration from vascular flora. It would be helpful if the extent of this bioclimatic influence could be determined. These desires could be satisfied by automatic weather recording stations located in the upper basin and close to the edge of Great Slave Lake. Alternatively theoretical modelling may provide an adequate representation of meso-climate relationships.

The good availability of net radiation data for Yellowknife Airport, (Wendland 1965, 1967), is marred by questionable representativeness owing to the omission of data for inclement weather periods. It might be expected that this data could be revised by a net radiation measurement program using the Airport site as a control in order that the original data could be corrected. The usefulness of this corrected data could be greatly extended if net radiation was also measured over other natural surfaces in the basin. Short period records spaced throughout the year could be related to the Airport control and adjustment factors determined for various surfaces at various times of the year. This line of research would thus utilise the principles set out by Garnier and Ohmura (1968) and Garnier (1972).

It should be noted that accurate net radiation measurement under conditions such as are found in the study area demands much effort and operator experience. Attending to the levelling, purging, checking and power needs of the measurement and recording devices is a full-time occupation if precise measurements are desired since such a level of attention minimises the inherent range of error in most net radiation measurement systems. From experience obtained in the present study it would be wise to ensure a good backup system or reserve instrumentation to deal with instrument malfunctioning problems.

It would be useful to run a ground heat flux monitoring program in conjunction with the net radiation program discussed above. It would be particularly desirable to study the vegetated depression sites such as muskegs and bogs, with a range of organic layer/mineral layer combinations (Nakano and Brown 1972). It is at such sites that the major potential for high summer evapotranspiration rates exist. If ground heat flux utilises a relatively high proportion of the available radiant energy the apparently low evapotranspiration rates through these covers becomes even more understandable. The peat, with its high moisture contents could function in the manner of a lake by storing heat which may be released when conditions are not so favourable for evaporation.

The potential importance of this factor in areas such as the present study area is indicated by the point made by Brown and Williams (1972, p.3) that the most extensive peat deposits in Canada occur in the discontinuous permafrost zone where conditions appear to be optimal for peat development.

Another useful program of research that could be run in conjunction with the net radiation studies is one focussing on wind and temperature gradient profiles over different surfaces. Such studies would aim at determining Bowen ratio values. It would be particularly useful to set up a transect of such profile stations from a bedrock-dominated surface through a muskeg or bog and out over a bordering lake. Very sensitive instrumentation would be required for this program. Wet and dry bulb temperatures have to be measured to an accuracy of 0.01°C. Sling psychrometers and ordinary mercury-in-bulb thermometers are much too insensitive for such work as was found out in an unsuccessful attempt to measure the gradients in the present study.

The combined energy balance and aerodynamic methods of determining evapotranspiration entailed by the above suggested research would also allow the exact significance of local advection energy in the study area to be evaluated. The physiographic mosaic that constitutes the Shield environment would appear to favour important local advection influences because of the high frequency of source-sink situations. The bedrock-dominated and vegetated surfaces appear to be likely sources of large amounts of advective energy and the lakes and other free water bodies the obvious sinks. The degree of significance of this effect would depend on the dimensions of the boundary layer zones in relation to the areal scale and mutual distribution of the units making up the Shield mosaic. This appears to be a fascinating area for future research which might interest a theoretical meteorologist since theoretical modelling could potentially provide much good evidence of likely intensities of the advective influence.

At the microscale it is possible that a battery of intermittent or small permanent pools could be set up and viewed as a black box with the aim of deriving information of a relative nature on advection influences, seepage patterns and general site evaporability.

At the macroscale the apparent importance of storage capacity changes on the water balance of individual years strongly suggests the need to instrument a set of lakes in the study area with water level recorders. Attempts could then be made to relate the patterns observed to runoff and precipitation patterns to enable better prediction of future water balance patterns.

This research has indicated the possible significance of long term storage changes as well as short cycle ones. The former changes could be researched further by biogeographical studies of selected lakes in the study area showing prominent vegetation rims around their borders. Dating of the peat or analysis of its contents and stratigraphic patterns may permit the identification of relatively wet or dry periods in the past. The levels associated with such periods could be surveyed and correlation attempted with other lakes. It may therefore be possible as a result of such studies to extend the hydrological record back in time and so enable better prediction of future patterns.

Further evapotranspiration studies would be useful in the basin. The lysimetric approach appears to be satisfactory particularly if the cans used are relatively deep and efforts are made to synchronise changes in the can environment with those in the natural surroundings. R.J.E. Brown is currently experimenting with a potentially useful lysimetric method at Thompson, Manitoba under conditions similar to those found in the present study area. Each lysimeter consists of a special half inch thick plastic pipe, eighteen inches in diameter and about fifteen inches deep. The bottom consists of half inch thick heavy plastic. A stilling well, made of lucite, is placed in each lysimeter near one edge. Thermocouples are installed at ¼ inch, two inches and four inches depth. The amount of water which needs to be added each day to bring the water level to a needle point in the stilling well, (in the manner of the Class 'A' pan operation), is recorded and added to any precipitation that has fallen to obtain a measure of evapotranspiration. The needle point in the stilling well is 4 inches below the top of the moss and is the result of an arbitrary suggestion by G.P. Williams, of N.R.C. (R.J.E. Brown; Telephone Communication, June 7, 1973). One risk associated with the lysimeters is the possibility of permafrost distrubance causing leaks in the containers if they are left in place through the winter.

The stilling well concept appears to be particularly useful at the permafrost sites. Similar wells could be planted in the natural surrounding area, in bottomless but filled lysimeters for instance, with the aim of obtaining information on the natural ground water level. Water could then be added to the lysimeters to maintain similar ground water levels there. If it is necessary to add water it is important that it should be done through an access tube down the side of the lysimeter rather than by sprinkling over the vegetation surface. The latter practice would produce artifically high estimates of evapotranspiration owing to the hygroscopicity of the moss and lichen covers. These covers would retain the sprinkled water which would evaporate rapidly. If the water is added from below, as is a more natural procedure for deep organic soils, it is probable that natural evaporation conditions would be obtained.

A more sophisticated means of attempting to ensure that the lysimeters simulate natural conditions is to install thermocouple psychrometers for the measurement of water potential (R.W. Brown, 1970). Such instrumentation would probably indicate the necessity of a drainage device for the lysimeter if used at sites like YT2 where relatively good drainage occurs in the latter half of the summer. This could be accomplished by siphoning methods in conjunction with a collection reservoir built into the base of the lysimeter at such sites.

Intensive soil moisture measurements could also be used, as in the present study to correct for limitations of the lysimeter. It would be wise to make such corrections for relatively short periods such as every three or four days.

Lysimeter studies could also be organised in a transect form along a soil moisture gradient from the edge of a lake through to surfaces where the bedrock is just deep enough to permit installation of the lysimeters.

It would also be desirable to have some check for the present study area for the generalisations made concerning transpiration rates through the higher order vascular flora. This could be accomplished by sophisticated soil moisture heat flux monitoring as suggested by Munn (1966, p. 39-41) and apparently utilised by Belotserkovskaya, Largin and Romanov (1967). The potometer method employed by Bliss (1960) may provide adequate results despite reservations stated about its applicability. It would be particularly useful to make comparisons for permafrost and non-permafrost sites and a variety of moisture conditions, in addition to relating any variations to the diurnal wind speed patterns present in the study area.

When suitable methods are developed, it would be useful to make diffusive resistance studies for different moss types to gain information on the possibility that conduction or sponge mechanisms are important substitutes for the transpiration process in this non-vascular flora.

Little consideration has been given to evaporation from snow in the present study. It has been assumed to be negligible but Landals (1971, p. 110) has indicated its potential importance during an extended snowmelt period for low albedo southfacing bedrock sites. The effect of its influence would be to further increase evaporation levels for such sites and to reduce runoff. There may therefore be worth in conducting snow evaporation studies in the basin during the snowmelt period using circular pans filled with snow, as suggested by Williams (1961).

The bedrock-dominated surfaces, the most predominant surface type in the area, were neglected in the field program for the present study. They are in fact worthy of much more detailed consideration from a hydrological point of view. Such studies are necessary to check the generalizations made here particularly on the interrelationship between storage capacities and the nature of the precipitation events. Perhaps, as Kakela (1973, p. 178) suggests a storm-oriented water budgeting approach would provide a suitable organizing model for these studies as opposed to a monthly - or even daily - oriented model like the Thornthwaite one (Bridge and Kakela, 1972).

All research conducted in Baker Creek basin must be tempered by the fact that the basin has suffered potentially significant disturbance by man either in the form of deliberate or accidental fires, in addition to natural ones (Rouse and Kershaw, 1971), or by chemical pollution. It would be most desirable to have an evaluation of this impact on the water balance. Chemical pollution appears to have substantially reduced the lichen cover in the basin in addition to killing or inhibiting the growth of many trees. It may be the case that the moss which has replaced the lichen functions in a relatively similar manner in evaporation terms. Also it is possible that tree growth was inhibited by natural conditions prior to the onset of chemical pollution. Thus little effective impact could be attributed to the phenomena. It is also possible that natural fires would negate the influence of deliberate or accidental fires caused by man's activities in the basin. Such hypotheses could be evaluated by research comparing the ecology of plots within the most affected parts of the basin and plots located in the same bioclimatic and physical belt but at some distance from Yellowknife. Potential locations may lie along the east and west roads out of Yellowknife.

The development of research in the Baker Creek Basin has lagged far behind expectations at the beginning of the International Hydrological Decade. In several years the project has been in danger folding completely. The suggestions for future research made above indicate the richness of research opportunities presented by the basin and the body of this thesis demonstrates the need to accept these opportunities. Much could be gained from a single broad-based and well-funded research program. This would be in contrast to the piecemeal, low-budget nature of the research that has been conducted in the basin. Such a program could fit in to a representative yet coarse and economically concentrated network of northern research basins designed to produce results for insertion into procedures which use readily available or readily inferred parameters (Garnier, 1972).

Baker Creek Basin should therefore be given strong consideration as an excellent location for future research aimed at determining patterns for a large portion of the central Mackenzie District, N.W.T.

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APPENDIX A

Taxonomy of Vegetation along transects made in close proximity to the main vegetative investigation sites. Two sets of transects were made each within about one mile of the Pocket Lake sites or the Muskeg Creek sites. The list presented here is an amalgam of that for each separate transect and it indicates the variety of *vascular* flora in this portion of the basin.

White spruce	(*Picea glauca*)
Black spruce	(*Picea mariana*)
Paper birch	(*Betula papyrifera*)
Water birch	(*Betula occidentalis*)
Dwarf birch	(*Betula glandulosa*)
Jack pine	(*Pinus banksiana*)
Aspen	(*Populus tremuloides*)
Larch	(*Larix larciana*)
Mountain alder	(*Alnus crispa*)
Speckled alder	(*Alnus rugosa*)
Common juniper	(*Juniperus communis*)
Creeping juniper	(*Juniperus horizontalis*)
Bog cranberry	(*Vaccinium vitis-idaea* var. *minum*)
Horsetail fern	(*Equisetum silvaticum*)
Horsetail fern	(*Equisetum arvense*)
Common fireweed	(*Epilobium augustifolium*)
Mountain fireweed	(*Epilobium latifolium*)

Dewberry	(*Rubus pubescens*)
Common bearberry	(*Arctostaphylos ura-ursi*)
Alpine bearberry	(*Arctostaphylos rubra*)
Labrador tea	(*Ledum groenlandicum*)
Labrador tea	(*Ledum palustre var decumbens*)
Prickly rose	(*Rosa acicularis*)
Lapland rhododendron	(*Rhododendron lapponicum*)
Mountain goldenrod	(*Solidago decumbens*)
Common saxifrage	(*Saxifraga bronchialis*)
Shrubby cinquefoil	(*Potentilla fruticosa*)
Wild gooseberry	(*Ribes oxyacanthoides*)
Black currant	(*Ribes hudsonianum*)
Crowberry	(*Empetrum nigrum*)
Blueberry	(*Vaccinium carspitosum*)
Arrowhead	(*Sasittaria cuneata*)
Cottongrass	(*Eriophorum augustifolium*)
Foxtail	(*Hordeum jubatum*)
Low bush cranberry	(*Viburnium edule*)
Cattail	(*Typha latifolia*)
Cloudberry	(*Rubus chamaemorus*)
Fern	(*Dryopteris fragran~~o~~s*)
Leather leaf	(*Chamaedaphne calyculata*)
Sweet gale	(*Myrica gale*)
Horsetail	(*Equisetum scirpoides*)

Northern Grass of Parnassum	(Parnassia palustris)
Willow	(Salix spp.)
Sedge	(Carex spp.)

APPENDIX B DAILY PRECIPITATION RECORD (inches).

APRIL-SEPTEMBER, 1966-1972.

YELLOWKNIFE AIRPORT, N.W.T.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	Total
1966																																
A					.53	.02	.01	.05	.04				.22	.02							T	T	T		T	T					/	0.87
M			.02	.11		T								T	.19				.01	T	T	.01	T	.01		T	T		.02	.29		0.66
J	.35	.11									T	.03			.22				T	.09	T	.31							.16	T	/	1.27
J						.14	.03				T	T				.08				.01	T	T	T			.19	.02	T		.12		0.59
A	.10	.01							T		.03	T		T	.02									T		T					.05	0.21
S	.15	.01		T	.05		.13	T	.04	.14	.09	T						.02	.01	T					T	T	.01	.14		.18	/	0.97
1967																																
A	.01		T		T	T						.08	T													T					/	0.09
M						.04		T		T			T	.04	T	T	.12	T	.01		.11	T			.01						T	0.33
J	.08					.01	.14	T		.03	.02								.01	.03					.08	.01	.06	.01	T	.31	/	0.81
J		T	.04	T				T	.24	.04		T										.03	.17		.02	.04		T	.05			0.63
A			.03			T					.03	.01	T	.08	.02					.01						.01	T				T	0.19
S	T	T	.01	T		.06	.01												.18			T					T	.20	T		/	0.46
1968																																
A						.02			.15	.23	.03							T	T				.04	T		T	T				/	0.47
M	.09	.08	.02	.05	T		T				.01	T	.24	T		T		.01	T						T	T				.02	T	0.52
J	.01	.05	T					T								.07			.01	.13	.08			T		.19	T				/	0.54
J	T				.11			T	.06						.01	T	T	.02	T	.10		.01	.07	.08	.01	T						0.47
A	.01										.01					.15	T				T			.08	.02		.01				.34	0.62
S	.72	.90	.02	T	T			.23	.01	.21			.06	T				T		T	.02		T	.03	.13	T		.04	T	.02	/	2.39

APPENDIX B (continued)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	Total
1969																																
A			.03	.06	T		.03	T					.05	T		T	.06	T	.09	.37	.03	.03	T								/	0·75
M					T			.11	.27		T	.19	T										T		.04	T	T			.23		0·84
J	T	T						.03							.04		.05		T	.01	T	.28	T			T	.01		T		/	0·42
J	T					.20	.01				.08		.11	.05							.01	T	T		.16	.01	.02		T	.01		0·66
A	.04			T		.13		T	T	.14	.02	.07		.02	.08			T	.38		1.78	.08	T		1·17	T	T	.14	1·08	.44	T	5·57
S						.01	.03			.06							.01						.04	.04	.01	.11	T			.25	/	0·57
1970																																
A	T	.01			.04						.02	.03	T						T	T	T		T	T	T	.14	.07	.01			/	0·32
M	.14	.08	T		T	.01									T	.17	.08									T	.02	.01		T		0·51
J	T	.01		T	T	.21				T	T	T			T			T	.02			T		.07	.11	.01	T				/	0·43
J			T	T	.22	T	.05	T	.04			.10	.01	T	.21	T						.10	.12		T	T					.54	1·39
A						T		T	.29	T	.97	.54	.03	T	.01	.01	.01	.63	T		.16	.10	.39	.06	.04	T	.03	.06				3·33
S	T			.02	.03	.07	.03	.10	.15	.03	.02			T	.03		.01	T		T	T	T		.02	.04		T	.12		.33	/	1·00
1971																																
A		.11	T			T	T								.06	.05	.05				.02	T		.01							/	0.30
M		.50			.05		.02	.14		T								.01	.01			.01		T	.03	T				.02	T	0·79
J								T	.14			.21		T														.14		T	/	0·49
J	.01	.02			.03	.13	.08	.04				.06		.01	.01	.04			.01	T			T	T					T	.01		0·45
A			.25		.32	.05		T	T	.03	.34	T	.04	T	T	.03	.69	.01		.12	.36	T	.16	T	T	T	.01					2·41
S				.04		T	T	T		.27	T	.08	.10	.13			.10	.05	T		.01	.05					T	T		.24	/	1·05
1972																																
A	·16				T	.13	.17							.09	T					.01					.04			T			/	0·60
M	·01			T									.13	T			.03					T									T	0·17
J	T		T									.03													T	.15	.22		.22		/	0·62
J	.16		.03	.03	.46		T	.14									.02	.07		T								.34	.22			1·47
A			.04				.14	.11	.10					T					.47	T					T	T	.01	.23	.13			1·23
S		·04	·02		.01			.01	.03	T	.03		.01		.72	.39	T	T							.15	.04	T		.01		/	1·46

APPENDIX C

GENERAL ENERGY BALANCE EQUATION AND THE BOWEN RATIO

ENERGY BALANCE: The principle of conservation of energy states that all gains and losses of energy at the surface of the earth must balance. The primary components are short-wave radiation from the sun, long-wave radiation from the earth and sky, transfer of heat through the ground, transfer of heat through the air, and the contribution of latent heat of evaporation or condensation. Other sources of energy can usually be neglected. The conservation principle can be expressed as a very general equation applicable at any instant in time.

$$\begin{aligned} Q_T - Q_R + Q_{L\downarrow} - Q_{L\uparrow} &= \pm Q_N \\ &= \pm Q_G \pm Q_H \pm Q_E \end{aligned}$$

where:

Q_T is the short-wave radiation from sun and sky, assumed by convention to be positive. It is absent at night.

Q_R is the short-wave radiation reflected from the earth. It is absent at night.

$Q_{L\downarrow}$ is the long-wave radiation received by the surface from the atmosphere.

$Q_{L\uparrow}$ is the long-wave radiation emitted by the surface.

Q_N is the net all-wave radiation. A gain of energy by the surface is positive.

Q_G is the transfer of heat through the ground. A downward flow of heat is positive.

Q_H is the turbulent transfer of sensible (ie. not involving changes of state) heat to the atmosphere. An upward flow is positive.

Q_E is the contribution of latent heat of evaporation and evapotranspiration. An upward flow of water vapour (evaporation) is positive; a downward flow (condensation) is negative. (Munn, 1966, p 3).

BOWEN RATIO: The Bowen Ratio (R) was introduced by I.S. Bowen in 1926 in an article entitled, "The ratio of heat losses by conduction and by evaporation from any water surface, (*Physical Review*, Volume 27, pages 779-789).

$$R = \frac{Q +1}{Q_E}$$

The value of the ratio may be estimated approximately from detailed data on wind, temperature and humidity gradients. Where measurements of net radiation and ground heat flux are available in addition to an estimate of the Bowen Ratio, it is possible to estimate the contribution of latent heat of evaporation and evapotranspiration from the equation:

$$Q_E = \frac{Q_N - Q_G}{1 + R}$$

(Terms are as defined above)

APPENDIX D

FIELD MEASUREMENTS

FROM EVAPORIMETER BATTERIES

N.B. The water loss value is the mean derived from all of the evaporimeters in each battery.

YT2:	Measurement Period	Precipitation (cms.)	Water Loss (cms.)
June:	2-7	--	1.42
	8-14	0.05	1.06
	15-21	--	0.65
	22-27	1.10	0.08
	28-7 (July)	1.52	1.47
July:	8-10	0.30	0.35
	11	--	0.09
	12-21	0.15	0.65
	22-23	--	0.13
	24	--	0.05
	25-29	1.30	0.29
	30-8 (August)	0.30	0.91
August:	9	0.49	0.06
	10-20	1.24	0.71
	21-23	--	0.66
	24-27	--	0.15

YT4:	Measurement Period	Precipitation (cms.)	Water Loss (cms.)
June:	2-7	--	1.01
	8-14	0.01	1.12
	15-21	--	1.28
	22-27	1.24	0.23
	28-7 (July)	1.65	1.39
July:	8-12	0.31	0.84
	13	--	0.21
	14-15	--	0.41
	16	--	0.17
	17	0.23	0.16
	18	--	0.22
	19-25	--	0.82
	26	--	0.17
	27	--	0.12
	28	--	0.09
	29	1.90	0.06
	30-3 (August)	0.17	0.78
August:	4	--	0.14
	5-10	0.52	0.48
	11-12	--	0.32
	13	--	0.14
	14	0.10	0.04
	15	--	0.07
	16	--	0.16
	17	--	0.13
	18-20	0.90	0.14
	21	--	0.19
	22	--	0.14
	23	--	0.20
	24-27	--	0.25

YT6M	Measurement Period	Precipitation (cms.)	Water Loss (cms.)
June:	7-14	0.10	1.66
	15-21	--	2.10
	22-27	1.30	1.00
	28-7 (July)	1.70	2.41
July:	8-12	0.30	1.05
	13	--	0.24
	14-15	--	0.44
	16	--	0.20
	17	0.18	0.14
	18	--	0.22
	19-25	--	0.67
	26	--	0.14
	27-28	--	0.06
	29	1.85	0.25
	30-3 (August)	0.30	0.74
August:	4	--	0.16
	5-10	0.51	0.41
	11-12	--	0.31
	13	--	0.08
	14-20	0.90	0.38
	21	--	0.20
	22	--	0.14
	23	--	0.17
	24-27	--	0.18

YT6S	Measurement Period	Precipitation (cms.)	Water Loss (cms.)
June:	21-27	1.30	0.77
	28-7 (July)	1.70	1.35
July:	8-12	0.30	1.86
	13	--	0.16
	14-15	--	0.30
	16	--	0.15
	17	0.18	0.12
	18	--	0.14
	19-25	--	0.56
	26	--	0.12
	27-28	--	0.08
	29	1.85	0.20
	30-3 (August)	0.28	0.14
August:	4	--	0.15
	5-10	0.51	0.24
	11-12	--	0.26
	13	--	0.09
	14-20	0.90	0.20
	21	--	0.05
	22	--	0.18
	23	--	0.12
	24-27	--	0.17